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ACTA GEOGRAPHICA 15 N:o 1

GLACIAL DRIFT IN ICELAND

ITS ORIGIN AND MORPHOLOGY

BY

VEIKKO OKKO

WITH 35 FIGURES AND ONE TABLE IN TEXT.
32 FIGURES IN PLATES I—XVI

HELSINKI 1956

Helsinki 1956. Valtioneuvoston kirjapaino

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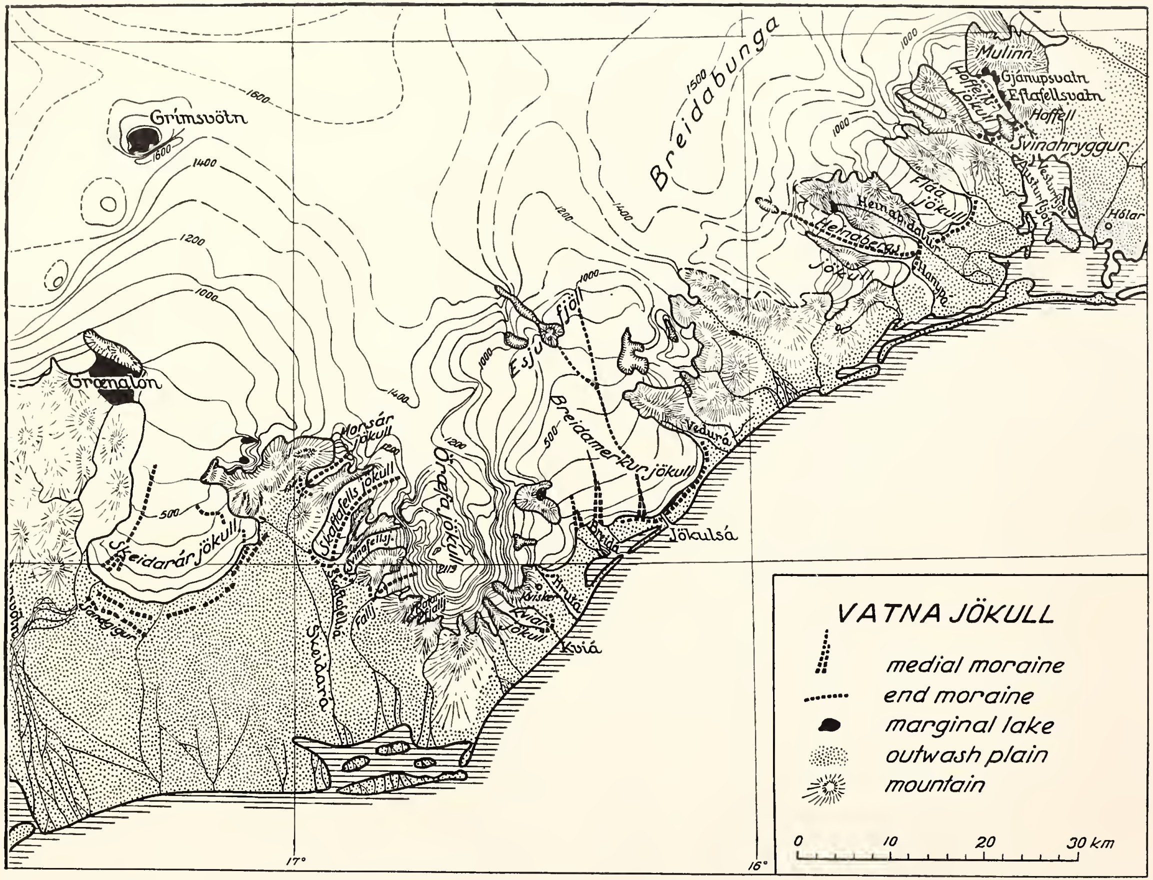

Fig. 1. The southern part of the glacier Vatnajökull. Compiled after various sources (Ahlmann 1937 c, Hjulström 1954 b, the topographical map sheets).

ACKNOWLEDGMENTS

In studying geography and geology at the University of Helsinki under the direction of my esteemed teachers Prof. Väinö Auer and Prof. Matti Sauramo, I learned to appreciate the importance of the work done by glaciers in building up the land surface and forming the landscapes of Finland. My understanding of the matter deepened upon my joining the staff of the Geological Survey of Finland, in the Department of Sedimentary Deposits, headed by Dr. Esa Hyyppä. I am profoundly grateful to my teachers for opening my eyes to see wonders created by Nature.

My knowledge of glacial morphology I was able to broaden on trips to northern Norway, Sweden and Denmark. In 1949 the Cultural Fund of Finland and the Emil Aaltonen Foundation awarded me grants to study the marginal zones of the existing glaciers of Iceland. I take this opportunity to express my appreciation. I set off on my journey in June of the same year and returned at the end of August. The main goal of my journey was the southern margin of Vatnajökull, but in addition I had an opportunity to acquaint myself also with the margin of Langjökull, terminating in the interior of the island, at Hagavatn.

I owe a debt of gratitude to my Icelandic friends, Sigurdur Thorarinsson, Tómas Tryggvason, Glúmur Björnsson and Helgi Jónasson frá Brennu, who have assisted me in many ways. Thanks are likewise due my splendid guides and hosts, notably Leifur Gudmundsson of Hoffel and Gudjón Jónsson of Flatey, whose contributions to the success of my trip are incalculable.

In the treatment and study of the material brought back with me from my trip, I have been assisted by the Misses Kyllikki Salminen, M. A., Kyllikki Parkkonen and Annikki Parkkonen, of the staff of the Geological Survey. My thanks to all three for work well done! Mrs. Lyyli Orasmaa and Mr. Erkki Halme have lent a hand in illustrating this study: the former has painstakingly done the drawings and the latter made the photographic prints. Some of the photographs are from Mr. Helgi Jónasson's extensive collection. Those who have helped me in my work I have to thank for much.

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Mr. Paul Sjöblom did the English translation under pressure to meet the publication deadline, for which I want to make known my appreciation.

Finally, I wish to thank the Emil Aaltonen Foundation for a second grant, with which it helped to carry this study through.

Helsinki, November 1955

Veikko Okko

INTRODUCTION

As a result of the effective glaciological research carried out in recent decades, it has become clear that the glaciers of today are geographically »living» entities, whose ice masses are continually being regenerated. From the glacier's *firn* or *accumulation area* the ice flows to the *ablation area*, bringing new ice to take the place of melting ice. If the annual accumulation and ablation are of equal magnitude, the *regimen* of the glacier remains in *balance;* but if the ablation exceeds the accumulation, the balance becomes negative. The regimen of a glacier is naturally reflected also in its other activity, such as the velocity of the glacier movement and the position of the glacier margin. Inasmuch as these factors affect the erosion, transport and deposition action of the glacier, its regimen must also be evident in the stratigraphy and forms of the glacial accumulations resulting from this action. On the other hand, it is to be expected that local conditions as well, such as the topography and quality of the glacier floor, add special features to glacial action, which emphasize the individual geomorphological behavior of each glacier.

The material transported by glaciers, considered in general, is herein termed *glacial drift*. According to the terminology of R. F. Flint, which the present author aspires to adhere to, »glacial drift embraces all rock material in transport by glacier ice, all deposits made by glacier ice, and all deposits predominantly of glacier origin made in the sea or in bodies of glacial meltwater, whether rafted in icebergs or transported in the water itself» (Flint 1947 p. 102). In addition to mineral matter, the present author includes under the term of glacial drift also the organic matter in transport by glacier ice, although its share, alongside that of mineral matter, is slight.

Thus understood glacial drift includes two parts: 1- the *load* being carried by the ice, and 2- the *deposits* formed therefrom but already freed from the ice. Load travels toward the glacier margin *superglacially*, *englacially* and *subglacially*, and its position in the ice varies during transport. Already during the transport stage glacial drift forms morphological accumulations on the glacier surface, e. g. surface moraines, which sometimes are called *active moraines* to distinguish them from *inactive moraines* freed from the ice. Part of the glacial drift is subject already in the moving glacier to *glacifluvial action.* This treatment takes place both on the glacier surface and, especially, under *subsurficial* conditions. It is often difficult to decide whether the subsurficial meltwaters flow along englacial or subglacial channels to the glacier margin. Upon being subject to the action of flowing water, glacial drift freed from ice is washed and sorted. It often forms its own accumulations, *glacifluvial deposits;* but sometimes glacifluvial material is contained quite abundantly in accumulations of glacial drift, or *moraines,* created by the direct action of the glacier. The prevailing material of moraines is usually, however, unsorted, unwashed and firm *till.* Inasmuch as the division of the material does not follow the morphological grouping of the formations, the present author joins Flint (1947) in regarding moraines as the morphological accumulations of glacial drift, alongside which glacifluvial deposits constitute another mophological accumulation category.

The treatment of glacial drift and the forms coming about as a result of it are at present under way in the marginal zones of the glaciers. In this respect Iceland is one of the best investigation areas. Glacial drift travels along with the ice to the glacier margin, where it becomes detached from the ice and forms various accumulations of drift, which numerous investigators have described. At the same time, the present formations of glacial drift in Iceland have been compared with those occurring in old glaciated areas. Already the classical studies of Torell (1857), Helland (1882), Keilhack (1883), Ussing (1903) and Thoroddsen (1905—06), brought the *sandurs* and *end moraines* of Iceland into scientific literature; and their counterparts were discovered in e. g. the glaciation area of Scandinavia. Later the marginal formations of Iceland were compared with e. g. the Salpausselkä ridges of Finland (Leiviskä 1928) as well as the marginal formations of North Germany (Woldstedt 1937, 1939). The glaciological features of the Icelandic glaciers did not become known, however, until the Swedish-Icelandic expedition published its investigations (Ahlmann and Thorarinsson 1937, 1938, 1939, 1940, 1943). It is against this glaciological background that the glacial drift collecting on the major glaciers of Iceland will be scrutinized in the following, including its transportation, its treatment at the glacier margin and its various accumulations. The material, structure and morphology of drift accumulations are placed in a causal relation to the general action of glaciers and from this basis a genetic grouping of the formations is arrived at. Finally the observations and conclusions made are compared with the results obtained in the study of other glaciers.

There is reason to underscore the fact, however, that the glaciers of Iceland are in a stage of rapid melting (Thorarinsson 1940) and the detailed features of their margins are constantly changing. As the observations presented in the following date back to 1949, it is probable that certain of the places described are by now different in appearance from what they were then, although no changes have taken place in the behavior of the glaciers.

THE EXISTING GLACIERS OF ICELAND

THE CONDITIONS OF GLACIERIZATION

The island of Iceland is one of Europe's most glaciated areas (Fig. 2). The island is built up of young volcanic rocks and their erosion products (Thoroddsen 1905—06, Pjeturss 1910). These rocks are generally soft, easily erodible, and they disintegrate from the effect of mechanical weathering. Consequently, exogenetic processes have left visible traces on the relief of the island. Glaciers and rivers have eroded valleys running toward the coast and along this coast there has evolved at an elevation of about 100 m a coastal plain, or *strand flat*, terminating in a precipitous cliff (Hjulström 1954). It continues under the sea to a depth of about 200 m,

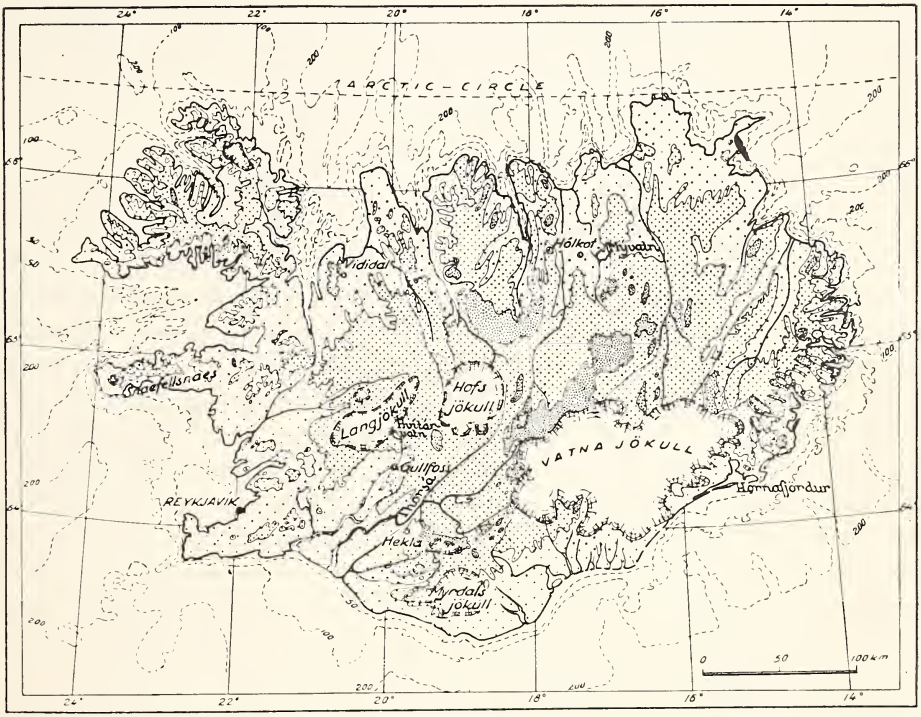

Fig. 2. The island of Iceland with its relief in rough and largest glaciers. Compiled from various sources.

as is to be noted e. g. in submarine contours. The strand flat (as the coastal plain is termed by Scandinavian geographers) is broadest along the southern and southwestern coast, whereas along the steep northeastern and northern coast it is narrow and broken.

The interior consists of a highland, varying in relief, with an elevation of 400—800 m above sea level. It is built of lava and tuff beds and their erosion products, and its surface is broken by tectonic fractures, which are clearly to be seen in the plantless highland landscape. A third elevation zone rising above the highland, consists of volcano cones and plateau horsts. Such eminences are scarce in the northern and central parts of Iceland, but in the southern part of the island they increase in number to the extent of forming a continuous mountain group beyond the strand flat, the tallest peak of which is over 2 100 m high. This mountainous barrier blocks the moist winds blowing from the Atlantic and forces them to rise. At the same time the moisture carried by them condenses and falls on the high southern part of the island in the form of snow. The interior and northern parts of Iceland, on the sheltered side, receives much less precipitation, for the north winds blowing from the Arctic Ocean seldom reach the island because the southwestern weather front usually extends over Iceland. The annual precipitation of the interior averages only 500 mm, as compared to 870 mm at Reykjavík (Thorkelsson 1946 p. 7) and 1 300 mm along the southern strand flat. On the northern side of the latter area an annual precipitation of as much as 3 000 mm has been measured on the glacier (Ahlmann 1939 a p. 50). Inasmuch as the greater part of the precipitation takes place in the winter months (Thorkelsson op. cit.), snow accounts for most of the total in the mountains and the interior. A continued accumulation of snow is possible only on the slopes of volcanoes and plateau horsts, which extend above the climatic snowline. This snowline was situated in 1936 in southern Iceland at an elevation of about 1 100 m, from which it rose to its maximum of 1 450 m in central Iceland. From there the snowline descended northward, being at about 1 000 m in North Iceland; and its lowest elevation of 700 m it reached in the northwestern part of the island (Ahlmann 1937 c p. 228). As above the snowline remain scattered and relatively small areas, which furthermore are situated far apart, the territory covered by ice is divided into several different glaciers.

THE ICE CAPS OF VATNAJÖKULL AND LANGJÖKULL

The present glaciers of Iceland cover an area of approximately 11 100 km², or 11.5 % of the total area of the island (Thorarinsson 1943 p. 18). Many of them are small valley glaciers on the northern slopes of mountains.

In addition, there are a few ice caps on the island characterized by a central dome formed on an elevated floor. Surrounded by a lobate margin, this dome either undulates gently or then rises evenly toward the center from every side (Ahlmann 1937 b p. 203). The largest ice caps are Vatnajökull, Hofsjökull and Langjökull, of which the firstmentioned is situated in the southern part of the island and the other two in the interior (Fig. 2 p. 9).

Vatnajökull comprises over 70 % of the glacierization area of Iceland and it covers 8 410 km² (Thorarinsson 1943 p. 17). Measured on the map (Fig. 2) the length of the glacier W—E is about 150 km and N—S about 100 km. The map (Fig. 1 p. 4) shows the highest part of the glacier to extend about 2 000 meters at the southernmost end of Vatnajökull, where the main ice dome connects with Öræfajökull, covering the upper slope of Iceland's highest mountain. The central ice plateau of Vatnajökull, extending across its northern end, is situated 1 300—1 800 m above sea level. Its highest points consist of two centers, of which the eastern one reaches an elevation of 1 400 and the northwestern one 1 800 meters. These centers are separated by a low, broad depression oriented N—S. At the southern end of the ice plateaus the relief becomes irregular and numerous nunataks pierce the surface. The topography of the floor divides Vatnajökull into several connected parts, each of which has its own outlets. The largest of Vatnajökull's subsidiary glaciers projecting southward are the following (Thorarinsson 1943 p. 17): Skeidarárjökull 1 722 km², Breidamerkurjökull 1 266.5 km², Heinabergsjökull 274 km², Fláajökull 244.5 km², Hoffelsjökull 312.5 km². Their outlets take the form of long ice tongues extending from the central plateau to the strand flat and their termini spread out in the form of round-curved expanded feet. The best developed are the expanded feet of the largest outlet glaciers of Skeidarárjökull and Breidamerkurjökull, the length of the margins of which are about 30 km. The profile drawn on the basis of seismic soundings (Fig. 3) gives a conception of the cross-section of Vatnajökull at the point of Breidamerkurjökull (Eythórsson 1953 p. 650). The expanded foot, according to this source, is 470 m thick at the center and its base reaches below sea level. The ice dome situated be-

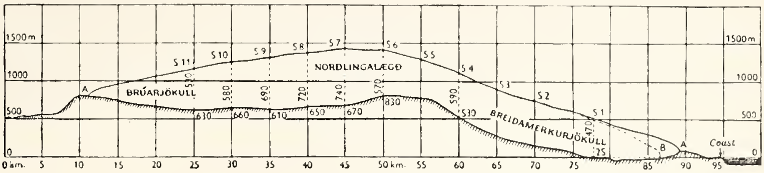

Fig. 3. A cross-section of Vatnajökull through Breidamerkurjökull, based on seismic soundings. The glacier surface is drawn after the topographical maps, while the dashed line shows the position of Breidamerkurjökull in 1951. By courtesy of J. Eythórsson (1953).

hind the outlet rests on a fairly flat base 800—900 m above sea level. The average thickness of the ice sheet there is 500—700 m, the extreme values at the measurement points being 1 014 and 360 m.

Between the five major outlets there remain numerous ice tongues resembling valley glaciers (e. g. Morsárjökull, Skaftafellsjökull and Kviárjökull), which terminate either at the edge of the strand flat or the valley hollowed out of the highland slope. Similar tongues of ice project outward also from the eastern margin of Vatnajökull, forming a »*Zungenrand*» (Spethmann 1912). Only in the north and in the west does the glacier margin rest on a relatively even floor. The ice tongues grow shorter there, taking the form of broad, curved lobes, which terminate in the highland already at 700 m above sea level. Spethmann (op. cit.) uses the term »*Lappenrand*». Simultaneously, the glacier receives a regular form: its surface is steepest in the marginal zone, from which the gradient decreases toward the center of the ice cap.

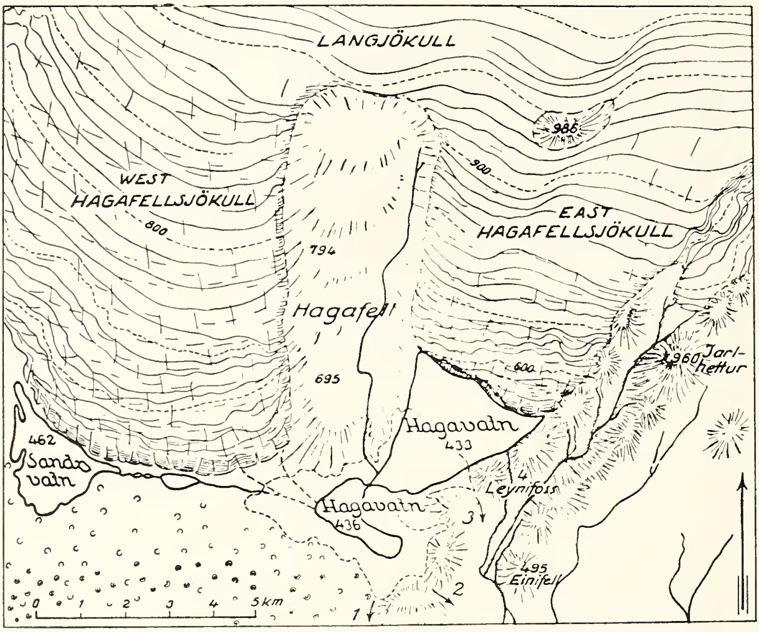

Fig. 4. Hagafellsjökull and Hagavatn at the southern end of the glacier Langjökull. The sketch is drawn according to the topographical map sheet No. 46, but modified after various sources (Wright 1935, Thorarinsson 1939 c). The dashed line around lake Hagavatn indicates the shoreline before the outburst in 1939. The outlets of the lake are indicated by arrows.

The ice mass of Langjökull (Thoroddsen 1905—1906, Bárdarson 1934) fills a valley about 75 km long and oriented NE—SW in the western half of the interior of Iceland. There are a few local glaciations in the mountains, over 1 000 m high, around this valley, of which Geitafellsjökull, attached to the SW part of Langjökull, extends about 100 m above the surface of the main ice dome, or to an elevation of 1 400 m. In the firn area of the main glacier there are two ice domes, the regular shape of which and the lack of nunataks indicate either thick ice or an even base, or both.

The ice tongues of Langjökull are generally short and broad and terminate in the highland at an elevation of 900 m above sea level. Only at a few points do they resemble broad valley glaciers. One of them extends as a twin projection into Lake Hvitárvatn, the surface of which is 419 m above sea level. At about the same level there terminates also the bi-lobal Hagafellsjökull, projecting out of the southern margin of Langjökull (Fig. 4). Its western lobe forms a round-edged arc in the highland valley, while the eastern lobe terminates in Lake Hagavatn, situated at an elevation of 433 m.

Generally speaking, the outlet glaciers of Langjökull are shorter than those of the southern margin of Vatnajökull and terminate at a higher elevation. They resemble most nearly the broad lobes at the northern margin of Vatnajökull.

THE REGIMEN OF THE GLACIERS

The regimen of Vatnajökull is known largely on the basis of the investigations carried out by the Swedish-Icelandic expedition in the eastern part of this glacier in 1936—38 (Ahlmann 1937, 1938, 1939, 1940 and Thorarinsson 1937, 1938 b, 1939 and 1943). It was proved that Vatnajökull ranks among extremely active glaciers.

Its activeness is exhibited in its heavy accumulation of snow above the climatic snowline, rapid melting below this line and swift ice flow from the central area to the margins. In the southern parts of Hoffelsjökull the climatic snowline is situated about 1 050 m above sea level. It divides the glacier in such a way that the accumulation area is 1.7 times as large as the ablation area. Though the accumulation area annually collects snow in abundance, the wastage in the ablation area exceeds even this intake. In the budget years 1936—37, 1937—38, 1938—39 the annual loss averaged 94 million m^3 of water. The glacier thus has a negative bilance. As a result, the glacier is at present shrinking. The wastage of the ice takes place chiefly as surface thinning, for the glacier margin has retreated on e. g. Hoffelsjökull relatively little in recent decades, though the glacier is known to have simultaneously become appreciably thinner. The annual ablation reaches a maximum (over 1 000 cm/annum) on the snout of the outlet terminating at an elevation below 100 m, but the wastage of ice is also promoted by calving of the glacier in the ice-dammed lakes along the sides of the outlet (Fig. 5).

The heavy accumulation causes a rapid movement of the ice from above the snowline toward the glacier margins. The effect of the glacier motion begins to be evident already in the accumulation area, where at the starting points of the outlets broad, spoon-shaped depressions are created in the surface of the ice cap. Toward the intake of the ice tongue the velocity of the motion increases and transversal crevasses open up on the surface; farther downstream these crevasses close. The glacier flow has proved to be most rapid at the neck of the outlet, which on Hoffelsjökull is situated about 500 m above sea level and about 7.5 km from the terminus.

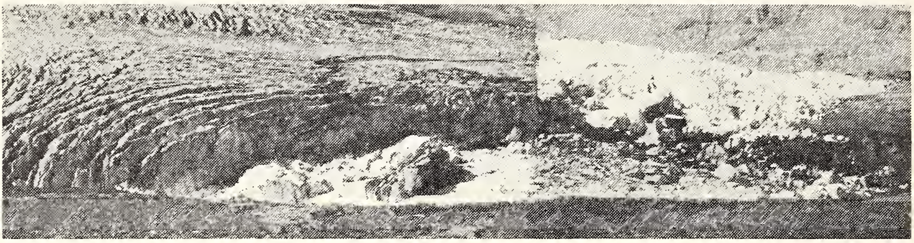

Fig. 5. Ice-dammed lake Gjánupsvatn after its drainage. The ice barrier (left) is broken and ice blocks fill the lake bottom. The black stripe in the ice (center left) is a volcanic ash layer. The ice wall is 10—15 m high.

Already at this point the surface of the outlet has become convex and since the transversal profile of its base is evidently concave, the ice body is thickest around its central line, where its surface flow is likewise most rapid. The maximum velocity of the surface of Hoffelsjökull, as measured in 1936, was 1 770 m/annum, but the velocity varies from year to year. E. g. the maximum velocity of the surface layer of the glacier measured in the summer of 1937 was 211 and in the summer of 1938 only 138 cm/day. The differences in the velocity of flow derived from fluctuations in accumulations, inasmuch as in 1938 the total was 30 % less than the year before. If the annual excess in certain years falls below average, the pressure is reduced and the glacier flow becomes slower. An above-average accumulation, again, makes the glacier flow faster and possibly causes an advance of the ice front.

Below the neck the flow of the glacier begins to slow down. This does not, however, take place evenly, but the velocity of flow depends also on the slope of the base. Downhill the ice advances so rapidly that its surface cracks or ruptures into ice cataracts. Below the drop the ice grows solid again. According to Thorarinsson's measurements the surface layer of

Hoffelsjökull moves at an average velocity of 15 m/annum 40 m from the snout. The terminus has thus become almost stagnant.

The regimen of the other parts of Vatnajökull to the south correspond to that of Hoffelsjökull. This is indicated e. g. by the fact that regardless of differences in shape and size there prevails on these subsidiary glaciers nearly the same glacial ratio between the ablation and accumulation areas. This ratio is 1 : 1.7 at Hoffelsjökull, Heinabergsjökull and Breidamerkurjökull, and 1 : 1.8 at Fláajökull; but it changes at Skeidarárjökull to 1 : 2.4 (Thorarinsson 1939 b p.196). Thorarinsson assumes that volcanogenic glacier outbursts account for the deviation (op. cit. p. 194). Inasmuch as these subsidiary glaciers are situated in climatically identical areas, their regimen and activity correspond. The subsidiary glaciers are in other words exceedingly active and their margins tend to move readily, as the results of measurements carried out in recent years show (Eythórsson 1949 b).

The regimen of Langjökull is not so well known as that of Vatnajökull. The climatic snowline there is relatively high, in the southeastern part of Langjökull being at an elevation of about 1 175 m (Ahlmann 1937 c p. 218). Judging by this, together with the shape and size of the ice tongues, the annual accumulation on Langjökull is smaller per unit area than in the southern part of Vatnajökull. As the pressure pushing the ice outward is thus correspondingly smaller, the ablation area is narrow and the ice tongues are shorter. The gradient of the glacier margin is thus great and stagnant ice occurs in front of the »living» margin only seldon. It began to form in the 1930's at the sides of the outlets terminating in Hvitárvatn (Nørvang 1937 p. 186). These outlets terminated in the lake as steep ice cliffs, and the glacier wasted away at these points by calving. The same kind of ice cliff occurs also at the terminus of the outlet extending to Hagavatn. This ice tongue earlier extended farther into the lake, covering the sills which divide the lake base. As the glacier margin retreated, one sill after another became exposed from under the ice and the outlet of the lake has shifted several times (Reynolds 1930, Wright 1935, Thorarinsson 1939 b).

In the behavior of the glaciers of Iceland there occur from time to time anomalous changes resulting from the volcanism. The best known of these are the glacier outbursts (»*jökullhlaup*» in Icelandic). They occur as sudden, heavy floods in the rivers fed by the glaciers. The cause of the floods is subglacial eruptions, which melt the ice around the craters. Meltwater forces a way under the ice and bursts forth through openings along the glacier margin, but at the height of a flood the water is discharged in a broad front from under the margin. The margin is thereupon shattered, forming ice walls scores of meters high. Glacier outbursts occur especially on Skeidarárjökull, on the ice plateau of which are volcanic craters (Nielsen 1937, Thorarinsson 1938 a). During the eight-day glacier outburst of 1934 about ten million m^3 of ice broke off the margin of Skeidarárjökull (Thorarinsson op. cit. p. 493) and at the same time 10—15 km^3 of water was discharged by the glacier (Nielsen op. cit.). This glacier outburst was of medium magnitude, according to Thorarinsson's estimate, for most of the glacier outbursts along the southern coast of Iceland wear away the ice the same way. The glacier outbursts recur at such long intervals, however, that the tunnels disappear and the glacier margins recover from the damage in the meanwhile. The effects of glacier outbursts are visible longest on the strand flat, where the masses of gravel and icebergs torn off the glacier margins are carried.

Subglacial eruptions at times also cause a local advance of the glacier margin. This happens when the ice mass cracks from the force of the eruption and the marginal part becomes detached from the glacier proper and slides into an area freed from ice. There it melts in place, because the glacier no longer nourishes it. Such an advance of a glacier margin took place in 1890 on the northern side of Vatnajökull, where the margins of Bruárjökull and Eyabackajökull advanced several kilometers. The glacier thereupon broke up into immense blocks of ice measuring as high as 200 m and at the same time a vast crevasse oriented W—E opened up (Thorarinsson 1938 a p. 495).

The frequent earthquakes occurring in Iceland have not been observed to cause changes in the behavior of the glaciers (Thorarinsson op. cit.). Evidently the thick ice constitutes such a plastic mass that the crevasses possibly opening up during a quake are quickly pressed together again. The ice does not crack at all during minor earthquakes (Nielsen 1937 p. 53). Nor have hot springs been observed to affect appreciably the activity of glaciers either, for the springs cause only a relatively slight rise in temperature at the surface of the earth. Locally, subglacial springs add to the volume of water discharged from the glaciers. As the water from the springs is often hot, it also melts the ice and thereby increases the quantity of meltwater.

Anomalous changes in the activity of glaciers are generally sudden and recur in the same places only after long intervals. Inasmuch as, moreover, they affect only a small part of the glacier at one time, the anomalous changes do not have any important effect on the regimen of the glaciers of Iceland. This rhythm depends on major climatic fluctuations.

As regards the earlier development of the glaciers of Iceland, it is known that the current negative balance has prevailed since the 1890's, when most of the glaciers had reached their maximum size during historical times and several of them perhaps during the entire postglacial period (Thorarinsson 1937 c p. 194, Ahlmann 1937 c p. 198). Before attaining this magnitude, the glacier margins had retreated even beyond their present position. This is indicated e. g. by the fact that the southern outlets of Vatnajökull yield fragments of peat and pieces and stumps of thick birch steams (Pjeturss 1907, Reck 1911, Bárdarson 1934, Ahlmann 1937). Bárdarson (op. cit.) places them on the level of the postglacial clima optimum, when the areas at present submerged under glacial ice would have been covered with birch forests and swamps. The same outlets also yield chunks of clay and shells of marine molluscs (Thoroddsen 1905—06, Pjeturss 1907, etc.). The mollusc fauna indicates an arctic sea of considerable depth (Pjeturss op. cit.). It may have extended up to the area currently covered by the outlets during either the late glacial period or Interglacial stages, of the latter of which there were either one (Pjeturss 1910) or two, (Nielsen and Noe-Nygaard 1936). The marine limit of the late glacial period is situated in southern Iceland at an elevation of over 100 m (Pjeturss 1901, Thoroddsen 1905—06, Áskelsson 1934, Kjartansson 1939). The shore-line is so weak, however, that its exact position has proved difficult to determine. Distinct shore cliffs have been found so far only at elevations of 40—50 and four meters above sea level (Bárdarson 1910). According to Bárdarson they are the product of marine transgressions, and to the lowermost, or 4 m shore, the sea rose from a level 2 m below the present water line. Thorarinsson finds a parallel to these stages in North Iceland, and in his opinion the 4 m shore corresponds to the younger Tapes-transgression in Norway, while the 40—50 m terraces are Gotiglacial. Iceland represents, according to this investigator, an upheavel area where isostatic reactions to changes in the size of the glaciers take place more rapidly than in Fennoskandia (1951). The Hólkot end moraines, situated some 50 km from the northern coast correspond in age, according to Thorarinsson, to the Salpausselkä-Raerne Stage in Fennoskandia (op. cit. p. 79). During the period of glacierization before it, there were probably ice-free areas on the Icelandic coast (Thorarinsson 1937 b p. 174), where plants were able to survive through the last Ice Age. The flora underwent a change during this time, however, for in the interglacial layers underlying the moraine there have been found e. g. *Alnus viridis* leaves, and among the pollen flora of the same layers there is an abundance of alder pollen (Áskelsson 1938, Lindal 1939). The most widely distributed of interglacial deposits consist of shelly clays, which have been met with e. g. at Reykjavík, Gullfoss and elsewhere in western Iceland (Pjeturss 1903, 1905, 1910, Bárdarson 1929 and Thorkelsson 1935). Their distribution indicates that during the Interglacial a large part of Iceland lay under the sea.

GLACIER TRANSPORT

The moving ice transports to the marginal zone various foreign material that in one way or another has become imbedded in the glacier or deposited on the surface of it. Part of this glacial drift consists of glacier erosion products while another part consists of foreign matter deposited on the surface. The latter includes wind-blown material, mechanical disintegration products, and in volcanic regions also ash layers.

The conditions prevailing in Iceland favor the forming of glacial drift. The glaciers are exceedingly active and the ice motion in them is fast. As under the glaciers there are soft and easily erodible species of rock and as the floor formed by them is furthermore in many places uneven, the quality and relief of the floor promote the erosive effect of the glaciers. Material is transported by the ice even more readily in places where it moves over loose deposits.

The conditions also favor the creation of drift on the surface of the ice. The often stormy winds blowing from the sea reach the glaciers freely, but first they must cross the land area surrounding the glaciers. In addition, the glaciers receive their share of the volcanic ash-falls; and, in the event of subglacial eruptions, coarse eruption products are likely to accumulate a top the ice in the vicinity of the eruptions. The uneven relief promotes also the formation of surface drift. From the effect of the relief the margin of the glacier in places receives a *Zungenrand* form and its projections lie tangent to the icebordering mountains. The latter, like the nunataks, are often precipitous. As in the climate of Iceland the mechanical disintegration of rocks is intensive, the surface of the rocks is covered with disintegration products, which slide down the slopes on top of the ice.

The drift participating in the glacier transport is divided by the author into wind-blown detritus, volcanic products, surface moraines, basal load, washed material and crevasse fillings. The classification has been carried out largely according to the manner in which the drift appears in the glacier, but at the same time it observes to a certain extent the grouping based on the origin of the material. The wind-blown detritus and volcanic products have come to the glacier from elsewhere and the glacier has only acted as the receiver of the material; the surface moraines contain in part disintegration products originating elsewhere, whereas the basal load has become formed out of the products of glacier erosion. The washed material includes mainly the washing products of the aforementioned and the crevasse fillings comprise the material accumulated in the glacier crevasses from the same.

In the following it will be shown how these drift materials of different origin occur in the glaciers and what happens to them during transport and what kind of accumulations they form on the ice.

WIND-BLOWN DETRITUS

Wind-blown detritus gathers on the glaciers of Iceland for the most part in the summer time, for then the winds are drier and the terrain around the glaciers is snowless. Moreover, the surface of the glaciers is damp in summer and this promotes the adhesion to the ice of eolian drift.

Eolian drift contains both minerogenic and organogenic detritus, which varies in amount, depending on the quality of the ground eroded. The content of organic matter in a sample taken from the Thorsá river valley in SW Iceland was 9.75 weight per cent, whereas the sample taken from Hoffel in SE Iceland contained 5.12 weight per cent.

Minerogene thus constitutes the main part of the eolian drift. It contains small fragments of volcanic rocks as well as mineral grains detached from the dried surfaces of fine-grained sediments. This material is blown by the wind in such abundance that e. g. the peat layers in bogs include minerogene to the extent of about fifty per cent. According to the studies of G. Jónsson (1930) the surface peat in the bogs of Iceland contains an average of 64.14 % minerogene, while the content in the peats of southern Iceland ranges between 42 and 53 per cent (Emilson 1911). During its transport by wind, the mineral material sorts itself. The prevailing fraction is fine sand, representing by a grain size between 0.5 and 0.02 mm (curves 1 and 2 in Fig. 6, also Emilson 1911, Hörner 1949), but strong winds also blow coarse sand and small pebbles, whereas light breezes blow off the ground only light organic waste matter, together with very fine sand and silt.

The organic component of eolian drift consists of the leaves of plants, pieces from twigs, stalks of grass, seeds, sedge fruits, and other debris. In addition, the organic matter includes pollen and spores from plants. The pollen rain originating from Iceland's present vegetation consists mostly of pollen from *Cyperaceae* and *Graminae*. Arborous pollen is represented by *Betula* pollen. This may be noted e. g. in two pollen diagrams (Figs. 7 and 8). The one (Fig. 7) has been made from the soil profile at Hoffel, SE-Iceland about 30 m above sea level and the other (Fig. 8) from the swamp at Laugardalur in the vicinity of Reykjavík, SE-Iceland (about 15 m above sea level). The pollen spectra of the uppermost samples in the diagrams represent the pollen rain of the present vegetation at the same time as the diagrams give an idea of the postglacial pollen flora. They show that the flora earlier contained *Betula* pollen in considerably greater abundance than nowadays. In the lowermost part of each diagram the *Betula* again declines and in its place there comes an abundance of *Ericaceae* and *Salix* pollen. Scattered occurrences in diagrams of *Alnus*, *Corylus*, *Pinus*, *Quercus*, *Tilia*, and *Ulmus* are to be interpreted as of distant origin as it has been shown by Aario in Finnish Lapland (1940, 1943). These trees do not belong to the

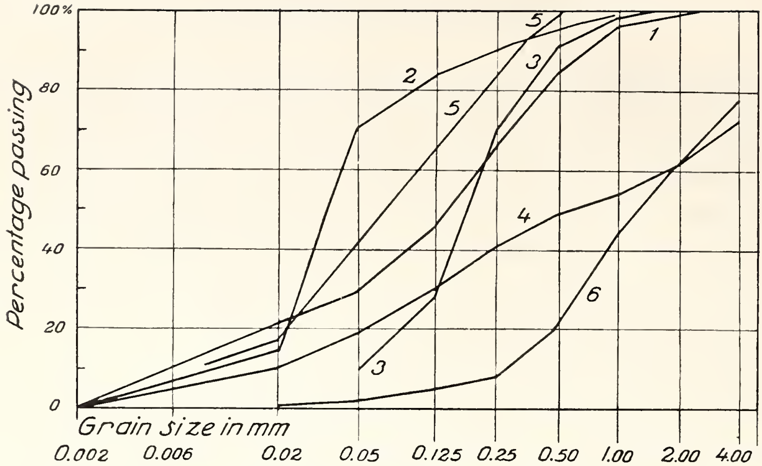

Fig. 6. Grain size distribution in the superglacial drift: Curve No. 1 eolian deposit from Hoffelsjökull, No. 2 the same from Breidamerkurjökull, No. 3 volcanic ash from Breidamerkurjökull, No. 4 basal load from a transversal moraine on Hoffelsjökull, No. 5 washed material from a delta on Hoffelsjökull, and No. 6 sorted drift from a lens in Hoffelsjökull.

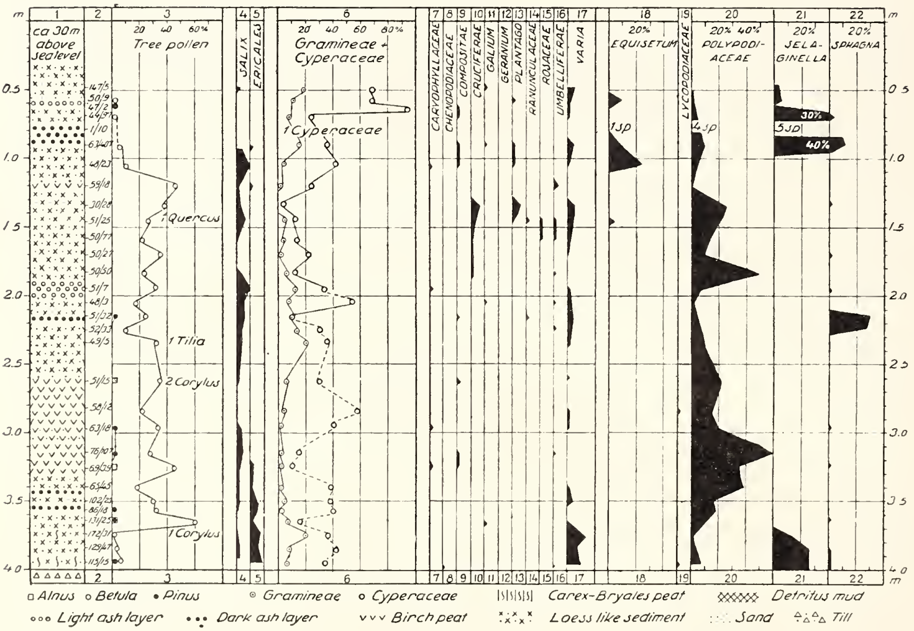

Fig. 7. Profile and analytical pollen diagram from Hoffel, SE-Iceland. The samples are taken from a section at the eastern bank of the sandur valley beneath Hoffels farm. The figures in column 2 indicate the amount of pollen and spores found in each sample. The percentages are counted from the common sum of pollen and spores. All samples are treated with HF.

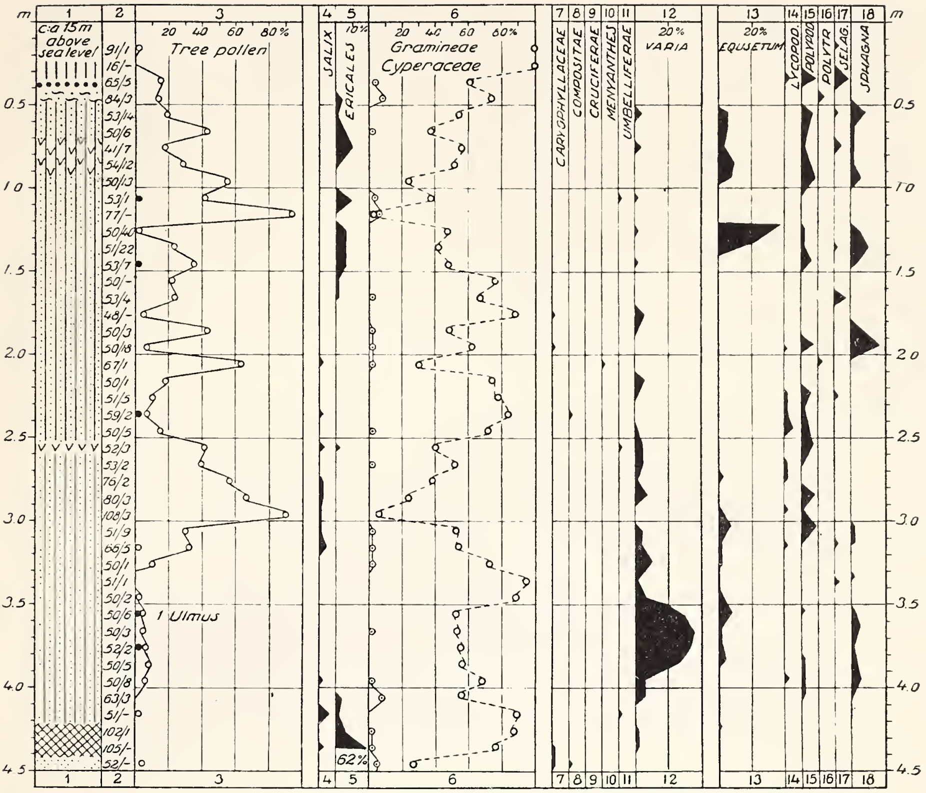

Fig. 8. Profile and analytical pollen diagram from a swamp at Laugardalur, near Reykjavík, SW-Iceland. The symbols used are the same as in Fig. 7. All samples are treated with HF.

present flora of Iceland. — Without undertaking in this connection any dating of the diagrams, we may note in their upper parts the same kind of regularity as in Thorarinsson's (1944) two pollen diagrams. The interglacial pollen flora of Iceland differs from this in that it contains an abundance of *Alnus* pollen. Áskelsson (1938) has investigated two samples from Snaefellsnes, W-Iceland, and Jessen two samples from Vididal, N-Iceland (Lindal 1939). The list of pollen and spores met with is following:

	Áskelsson		Jessen (Lindal)	
	Sample H	Sample K	Sample 1	Sample 2
Alnus	1 %	44 %	5 sp.	43 %
Betula	3 »	—	3 »	35 »
Pinus	4 »	—	—	—
Picea	1 »	—	—	—
Salix	10 »	3 »	—	22 »
Caryophyllaceae	3 »	—	—	100 %

Chenopodiaceae	4 %	—	—	—
Compositae	1 »	1 %	—	—
Cyperaceae	15 »	5 »	—	—
Ericales	18 »	—	—	5 sp.
Graminae	32 »	46 »	—	40 »
Hippophae ?	1 »	—	—	—
Plantago	1 »	—	—	—
Varia	6 »	1 »	—	6 »
	100 %	100 %		
Dryopteris			—	2 »
Lycopodium			1 sp.	1 »
Sphagnum			—	2 »

In the eolian sediments originating during the postglacial period there are naturally the pollen and spores of the same species of plants as appear in the diagrams presented in the foregoing, but their abundance is affected not only by the vegetation prevailing in the environment but also by the season of the formation of the sediment. If the eolian sediment has accumulated during the early summer it contains in the main pollen from plants blossoming at that season. Thus, the pollen spectra of eolian sediments vary according to the time of year the winds happen to be blowing. The most important observation, however, is that eolian drift continuously gathers on the glaciers of Iceland, with the drift containing not only waste matter but also recent pollen flora. E. g. a sample taken from the dirt cone of Breidamerkurjökull contained the following pollens:

Betula	5
Salix	1
Artemisia-type	1
Caryophyllaceae	1
Cyperaceae	2
Graminae	5
	15

In addition, 1 *Polypodiaceae* and 1 *Sphagnum* spore were met with.

Eolian drift accumulates practically everywhere on the glaciers of Iceland. It stands out from the white ice and snow by being dark and dirty. In the ablation zone it forms on the surface of the glacier a fairly uniform, if thin, film. The dirt layer is thickest in cracks and hollows in the ice. On level surfaces it fills rough spaces. In winter the surface dirt is likely to freeze onto the glacier, but the thaw of the following summer releases it anew. As new matter gathers on the surface of the glacier each year, blown by the wind and exposed by melting, the amount of surface dirt increases steadily. Simultaneously, the dirt layer moves along with the ice toward the glacier margin.

Besides the loose surface dirt there is to be observed in the ice of the ablation zone dark bands containing eolian dirt. They form with the cleaner ice alternating, almost parallel stripes, which arc in the direction of the glacier margin. Such dirt bands occur in the ablation zones of both the Vatnajökull and Langjökull glaciers. Whenever cracks cut the dirt bands and whenever the glacier margin has been ruptured and become steep, it may be noted (Fig. 9) in the vertical sections that the dirt bands continue into the ice. The bands, in other words, have formed in places where layers of ice containing dirt intersect the surface. These ice layers slope back downward, though their declivity varies. At Skaftafellsjökull the dip of the dirt bands was steepest, or 40—45°, about 200—300 m from the glacier snout; but it became more gentle both toward the end of the ice tongue and the snowline. The gentlest dips varied between 10 and 15° in both directions. A change in dip was observed also when measurements were carried out along the same dirt band. The gentlest positions were obtained in the region of the center line of the ice tongue, while the dip became steeper towards the sides of the ice tongue at the same time as the course of the band turned, following the margin, and finally ended up nearly at right angles to the course, as measured from the center line. Observations indicating the same tendency were also made at Hagafellsjökull.

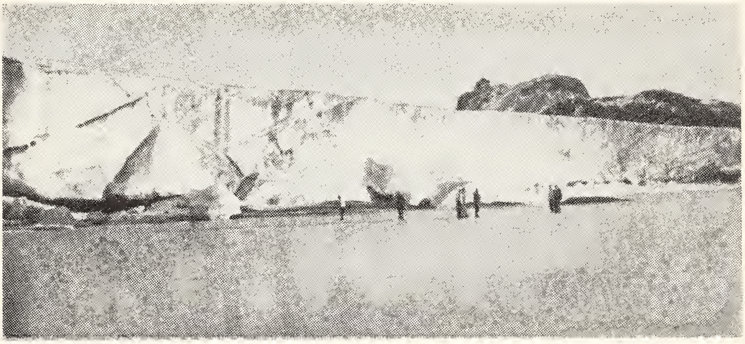

Fig. 9. The front of Hagafellsjökull forms a steep cliff in the lake of Hagavatn, which is covered by winter ice. Englacial drift appears as dark bands in the ice wall. The peak of the mountain Jarlhettur in the background. Photo Helgi Jónasson, 1951.

The strike and dip of the dirt bands on Hoffelsjökull proved to deviate from the foregoing. There, too, gently sloping dirt bands occurred in the glacier snout, oriented in the direction of the margin and inclined upstream. Their dip varied greatly, however, and the bands occurred brokenly. This was because the glacier had broken up into blocks of ice, which had evidently moved in relation to each other, causing breaks in the dirt bands. Deformation of the dirt bands was also noted at Morsárjökull, where the glacier breaks off in chunks from the top of a precipice, forming a heap of ice at its foot. As a result, the dirt bands that previously had run parallel are broken up and their dip turns every which way. Furthermore, new dirt (as well as winter snow) gathers in the openings between blocks of ice, adding to the diversity of the ice breccia.

The amount of eolian drift varies in each individual dirt band, just as on the surface of the glacier. At times the dirt occurs as such weak impregnations that it promotes melting of the ice so that the surface of the glacier becomes depressed where it contains dirt. Usually, however, there is so much dirt that the layer of it melted free of the ice retards the melting of the ice underneath. The band thereupon becomes prominent on the surface of the glacier as a sharp, dark, ridge-like formation. Such dirt ridges rose on the southern outlets of Vatnajökull a maximum of a meter above the glacier surface. The surface of the dirt ridges consists of a loose layer of drift melted out of the ice, and from it the dirt spreads over the ice surrounding the band.

Both in the Alps (Vareschi 1935, 1937, 1942) and in Alaska (Heusser 1954) pollen analyses have revealed that the substance of the dirt bands consists of eolian sediment embedded in the ice. Quite evidently, eolian matter is deposited in summer on the surface of the accumulation area and is then covered with snow, thereafter becoming buried in ice, Such layers are transported along with the ice to the ablation area, where they become exposed as bedded bands in the melting ice. The eolian sediment thus brought up anew to the surface and freed from the grip of the ice moves with it until the movement of the glacier stops. Simultaneously this surface drift »riding» on top of the ice promotes roughening of the surface by in places retarding and in other places accelerating the melting of the ice.

The same kind of movement of eolian drift occurs when dirt gathers in crevasses and depressions in the ice. The transversal crevasses occurring at the climatic snowline likewise collect dirt. When the crevasses close up, the accumulated dirt becomes embedded inside the ice and forms bands and lenses. Corresponding gathering of dirt also occurs in the crevasses of the ablation area. Only the dirt accumulating in the open radial crevasses of the glacier terminus no longer becomes embedded within the glacier.

The dirt cones (Fig. 1, Plate I) occurring on the eastern part of the glacier snout of Breidamerkur, the eastern margin of East Hagafellsjökull and Hoffelsjökull were conceivably created by the dirt collecting in cracks. Such cones are formed upon the melting of dirt lenses in the ice and are often situated apart from the dirt bands proper. When such a dirt lens comes to the surface as a consequence of melting, it prevents the ice underneath from melting, whereupon the area containing dirt becomes covered with a melted layer of dirt and there forms a sharp-pointed ice pyramid (Spethmann 1908, Wilson 1953).

VOLCANIC PRODUCTS

On the volcanic island of Iceland some 30 eruption centers are known that have been active during historical times. In addition there are countless ancient craters on the island. The number of eruptions exceeds the number of craters many times over, as the eruptions often occur in the same centers. E. g. Hekla is known to have erupted 23 times since Iceland was settled (Thorarinsson 1944). During an eruption the vicinity of the crater becomes covered with various eruption products. There is a rain of ejectments, especially during the initial stage on an eruption, when the atmosphere often becomes filled also with volcanic ash. It spreads out from the eruption center with air currents in the direction of prevailing winds, forming a fanlike pattern upon raining down, the tip of the extended fan being at the eruption center. As the winds in Iceland most generally blow from the southwest or south, most of the sectors of ash falls open to the northeast, north and east. The volcanic ash is sorted when blown by the wind in such a way that the coarsest and, in part, also the heaviest material falls closer to its source than the fine silicate dust (Thorarinsson op. cit. p. 87). The latter material has been observed to travel thousands of miles, for volcanic ash from Hekla has been blown as far as Finland (Salmi 1948). In the soil profiles of Iceland volcanic ash layers may be seen in the form of horizontal bands lying one on top of another. As volcanic activity occurs mostly in the southwestern part of the island, the western and central areas of Iceland most often receive ash falls. Moreover, the ash layers in those areas are thicker and contain coarser material than in the eastern part of the island. At Skallakot, which is situated about 20 km NW of Hekla. Thorarinsson has found 14 ash layers in the soil profile (op. cit. p. 50), whereas in the Hoffel profile (Fig. 7 p. 20) there are six ash layers and in the Laugardalur profile (Fig. 8 p. 21) there is only one.

Volcanic products accumulate also on glaciers. For the most part, they consist of sorted ashes, as coarse eruption products are deposited on the ice only in the vicinity of subglacial eruption centers. When Grimsvötn erupted in 1934, volcanic gravel was deposited, according to an estimate by Nielsen, to a thickness of 20—25 m in the vicinity of the eruption center in Skeidarárjökull. Since it was hot upon falling onto the ice and cooled down slowly, the gravel melted the ice underneath and remained snowless. Only after cooling down, the gravel was covered with snow and became embedded in the glacier (Nielsen 1937). During the same eruption volcanic ash fell also onto the eastern and central parts of Vatnajökull, but it formed on the ice a thinner and finer-grained layer, the ash of which had cooled down already during its transport through the air and was quickly buried under snow (Ahlmann and Thorarinsson 1939 a p. 39). At Hoffelsjökull Thorarinsson estimated that during the past 1 000 years only 5 cm of volcanic ash has fallen (Thorarinsson 1939 b p. 213).

Volcanic ash acts on the surface of a glacier like eolian drift. In the accumulation area it becomes buried under snow and imbedded within the ice, and in the ablation area it forms on the surface of the ice a loose, dark layer, which becomes subject to the action of rain and meltwaters. In ablation areas also layers of volcanic ash deposited in accumulation areas come into view again, when they form broad stripes on the surface of the ice, recalling dirt bands. The ash layers wind parallel to the glacier margin in dark belts, and they stand out already at a distance from the lighter surface of the ice between them. Such ash belts were photographed by Palmi Hannesson in 1938 at Skeidarárjökull (Thorarinsson 1944, p. 122). Similar dark ash belts were seen also in 1949 on the surface of both Skeidarárjökull and Breidamerkurjökull. These ash belts were transported slowly along with the ice toward the glacier margin. An ash belt forms on a glacier at a place where an ash layer buried in the ice meets the surface. Like dirt bands, these layers dip down backwards. In broken up parts of a glacier the ash layers are also deformed (Fig. 5 p. 14).

With the melting of the glacier surface the volcanic products become detached. The ash band extending up to the surface is thereupon covered with a loose ash layer, which retards the melting of the ice underneath so much that the ash belt is raised above the surrounding surface. There form crests and pyramids of ice covered by ash, remindful of the surface formations of eolian drift. From the steep slopes ash is washed into hollows on the glacier surface. At the same time the volcanic matter becomes mixed up with eolian and other surface drift. Thus e. g. the ash cone occurring on the surface of Breidamerkurjökull contained 0.24 % organic matter. The sorting and grain size of the ash are revealed in Fig. 6 p. 20.

The share of volcanic products in the drift varies in the glaciers of Iceland both in respect to time and place. They occur most abundantly in the surface drift of ablation areas when the layers formed in the accumulation areas reach the glacier margin, yielding there the material contained in them. Locally, volcanic products are mostly met with in the western part of Vatnajökull and there on the ice tongue of Skeidarárjökull. With the exception of the coarse eruption products heaped around subglacial eruption centers, the volcanic products accumulated on the glaciers of Iceland are sorted. The grain size ranges from coarse to fine sand. These deposits usually contain more coarse material than eolian drift, which they resemble in many ways.

SURFACE MORAINES

By surface moraines are herein meant accumulations of glacial drift occurring on the surface of a glacier that have a certain form and trend. Their material is for the most part unsorted. Surface moraines are classified as medial, lateral and transversal.

Looked at from the air, there can been seen on the southern outlets of Vatnajökull dark belts running downstream. These belts, at times winding and at times directed at right angles toward the ice margin, are medial moraines (Fig. 1 p. 4). According to the maps, medial moraines occur on most of the southern outlets of Vatnajökull. All the starting points of the medial moraines are situated below the elevation of 1 000 m, or below the climatic snowline. The majority of the medial moraines start from nunataks or come to the glacier surface immediately below them. Some medial moraines, however, such as those of Skeidarárjökull, Morsárjökull and certain ones of Breidamerkurjökull appear on the glacier at spots where outcropping rocks are lacking. Many medial moraines proceed in a fairly straight line from their point of departure to the glacier margin, but some (those of Skaftafellsjökull and Morsárjökull) arch out, following the outlet lengthwise, while others wind considerably (the easternmost medial moraine on Skeidarárjökull).

The situation of the medial moraine on the glacier surface determines the point of departure. If the starting point of a medial moraine is situated approximately on the center line of an ice tongue, the moraine runs along the center of the tongue to the glacier margin (the medial moraines of Morsárjökull, Breidamerkurjökull and Heinabergsjökull). If, again, the medial moraine begins at one side of the outlet glacier, it advances also along this side (e. g. the medial moraine of Skaftafellsjökull runs from the nunatak east of the center line, remaining just about equidistant between the lateral side of the glacier and its central line). The two largest medial moraines of Skeidarárjökull are situated in such a way that one runs near the western margin and the other near the eastern margin of the outlet. The former divides at its lower end into branches, spreading out fan-like apart from each other and terminating within the bounds of the glacier.

The medial moraine of Heinabergsjökull is likewise marked as terminating before reaching the glacier margin. In 1949 this medial moraine extended as far as the glacier margin (Fig. 10), but the margin at the time was 450 m farther back than is shown in the map (Eythórsson 1949 b). With these exceptions, the medial moraines carry their material up to the glacier margin. For most of the distance they are relatively narrow moraine belts and they spread out, like on open fan, only at the margin of an outlet.

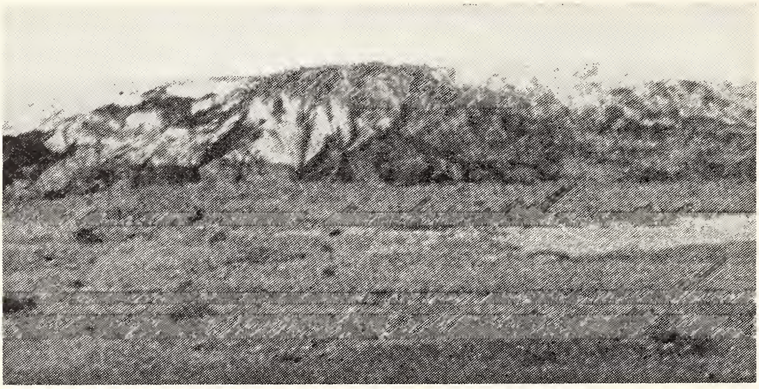

Fig. 10. Radially crevassed terminus of Heinabergsjökull bears a dark medial moraine on its surface. White clean ice comes into sight in crevasse walls. In front of the glacier an ablation moraine.

Thus the medial moraine, which is the longest in Iceland, running along the center of Breidamerkurjökull to the glacier margin, starts as two narrow branches from the nunataks of Esjufjöll. The branches join at an elevation of about 500 m, forming a moraine belt over 10 km long and 250—200 m broad. It begins to widen out about four km from the glacier margin, and a km away it is already 1.2 km broad. Toward the margin of the glacier the width of the medial moraine expands rapidly and it reaches the margin on a front approximately 2 km broad (Fig. 11).

At the same the glacier is crevassing. Radial crevasses are also thereupon created in the part covered by the medial moraine. Their walls stand out from the dark moraine-covered surface by virtue of their whiteness, for both at Breidamerkurjökull and Heinabergsjökull clear ice shows under the medial moraine. The same observation is to be made also at Hoffelsjökull, where the slight medial moraine derives from the Mulinn mountains at the eastern margin of the neck of the glacier. In the summer of 1949 numerous cracks ran across the drift-covered ice, and clear ice shone through from the walls of the crevasses. These observations indicate that the medial moraines originating below the snowline from nunataks and marginal mountains have formed on the glacier surface and been transported downstream by the ice. Ot the other hand, it is evident that the medial moraines originating from nunataks above the snowline as well as from all the peaks of the bedrock covered by the glacier are first embedded in the ice, where they are transported invisibly until exposed by melting. The medial moraines thereby created appear to grow straight out of the glacier.

The drift layer of medial moraines is generally so thick that it effectively retards the melting of the ice underneath. The moraine belt accordingly rises above the rest of the glacier surface and forms on the ice a long and narrow ridge. The slopes are steep, but the summit is comparatively flat. The glacial drift deposited in this moraine belt thus, as it were, rides on top of the glacier, moving at the same speed as the ice transporting it. During transportation the material is no longer subject to wear. This is evidenced e. g. by the fact that moss-covered stones have been found on the medial moraine of Vatnajökull — the vegetation, as it is explained, having developed during the transport phase (Eythórsson 1951, p. 503). The material of the medial moraine is unsorted and contains an abundance of boulders and stones. These are usually angular and sharp-edged (Fig. 2, Plate I), but glacier erosion marks, facets and striae are seldom observed on them. Petrographically they represent almost exclusively the same rock type, for the starting point of the same medial moraine ordinarily consists of a relatively small rock outcrop.

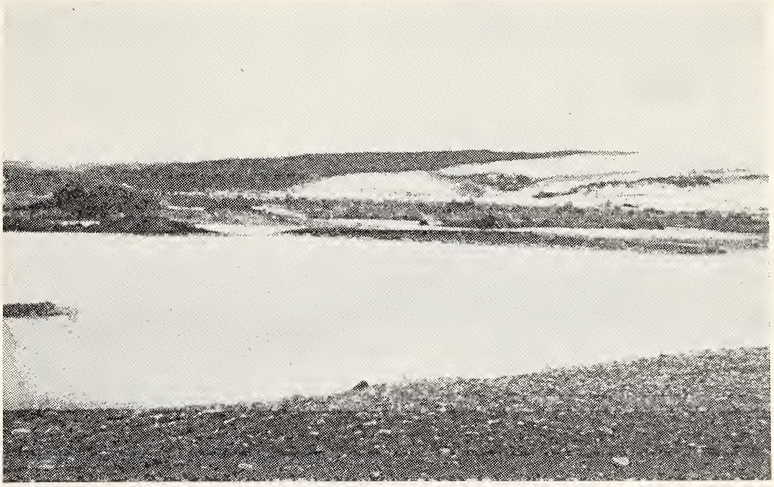

Fig. 11. The terminus of Breidamerkurjökull with its large medial moraine sinks down into the marginal lake at the source of the river Jökulsá.

The material of the medial moraine continues to represent the quality of the initial material even in the marginal zone. Especially its stones and blocks retain their original form during transport. If the point of origin of the medial moraine is situated underneath the glacier, it contains exclusively the products of glacier erosion. If, again, the point of origin of the medial moraine is a nunatak cropping out of the ice, the material of the medial moraine will include, in addition to erosion products, also mechanical disintegration products.

Lateral moraines originate at the line where the glacier margin meets the surrounding mountainsides. Frequently, they cover the side of the glacier as a narrow belt. Lateral moraines travel along with the glacier in the direction of the ice flow and often form the start of medial moraines. A lateral moraine receives a large part of its material in the form of frost wedging and disintegration products sliding down the valley slopes as well as loose deposits covering steep rock walls. This could be plainly observed e. g. at Skaftafellsjökull and Hagafellsjökull, at the joints of which there still remained in the summer of 1949 snow from the previous winter. On top of this snow stones had fallen down the rock walls, leaving tracks to mark their trail. As the snow underneath the lateral moraine thus formed was clean, it would seem that the avalanching of the material had taken place mainly during the summers. The material thus falling down nearer the accumulation area from steep mountainsides is evidently buried in wintry snow and becomes enveloped by the glacier. In the ablation zone, on the other hand, they accumulate on the surface of the ice sheet and there form an ever-thickening mantle, which covers the side of the ice tongue (Fig. 3,

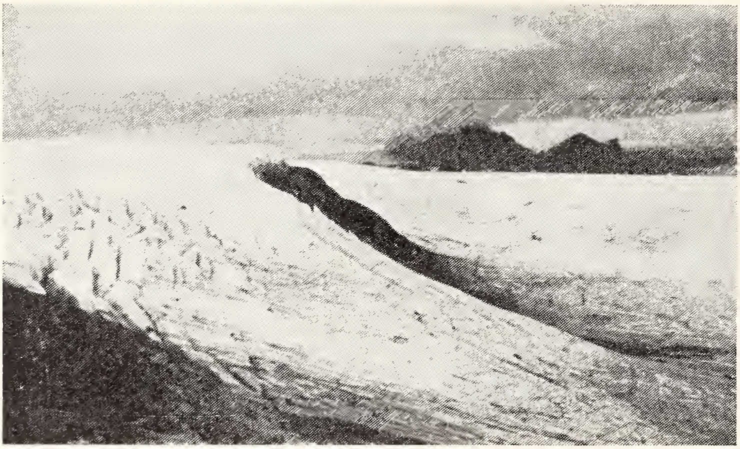

Fig. 12. A small nunatak piercing East Hagafellsjökull dispatches no medial moraine. Photo Helgi Jónasson, 1946.

Plate II). Such lateral moraines occur especially at the lower ends of outlets, where they grow to be several meters thick. E. g. on the western side of Skaftafellsjökull the lateral moraine was observed to cover the ice below as a layer more than two meters thick (Fig. 4, Plate II). As the melting of the glacier is faster beyond the lateral moraine than underneath it, there forms between the lateral moraine and the glacier on the surface of the ice a valley paralleling its edge, whence the lateral moraine becomes detached and forms on the mountain slope a bank containing ice. Before this can happen, the lateral moraine is likely to be transported by the side of the glacier and erode the joint between glacier and rock, with the result that its material becomes mixed with erosion products. The ice must thereupon extend in relative thickness to the mountain slopes. If, however, the edge of the glacier is thin and its movement slow where it adjoins a nunatak or lateral mountain, its transportation capacity is so slight that disintegration products will make up the main substance of of the lateral moraine. This conception is supported by the fact that on small nunataks newly projecting out of the ice and still lacking in disintegration products and to which the glacier extends as a thin sheet (Fig. 12) there occur no lateral moraines whatsoever and from them no medial moraines originate either.

From the foregoing it follows that the material of which lateral moraines are composed include an abundance of products of slumping and rolling. In addition, products of glacier erosion are also likely to the found. The amount of each component depends both on the erosion capacity of the the ice border and on the position and steepness of the rock walls surrounding the ice. As lateral moraines often act as the sources of the material composing medial moraines, it is understandable that a substantial part of the medial moraines thereby originating consists of the products of disintegration and frost wedging. For this reason, there is an abundance of sharp, jagged rock fragments in the material, on which erosion marks often are lacking.

In the ablation zones of Icelandic glaciers there occur also surface moraines whose accumulations are situated parallel to the glacier margin. Woldstedt (1954 p. 20) calls them transversal moraines *(Quermoränen)*. At Kviárjökull such a transversal moraine has formed on top of stagnant ice in front of the »living» glacier (Todtmann 1936). The transversal moraine occurring at Dyngjujökull in the northern part of Vatnajökull has evolved in the same way (Spethmann 1909 p. 422, Woldstedt 1937). Both are extraordinarily large, for ordinarily transversal moraines occur as a number of successive bands parallel to the ice margin. Moraine bands have formed in places where glacial drift, filling forwardrising transversal crevasses, meets the ice surface. Often the drift forms a wall-like ridge along the

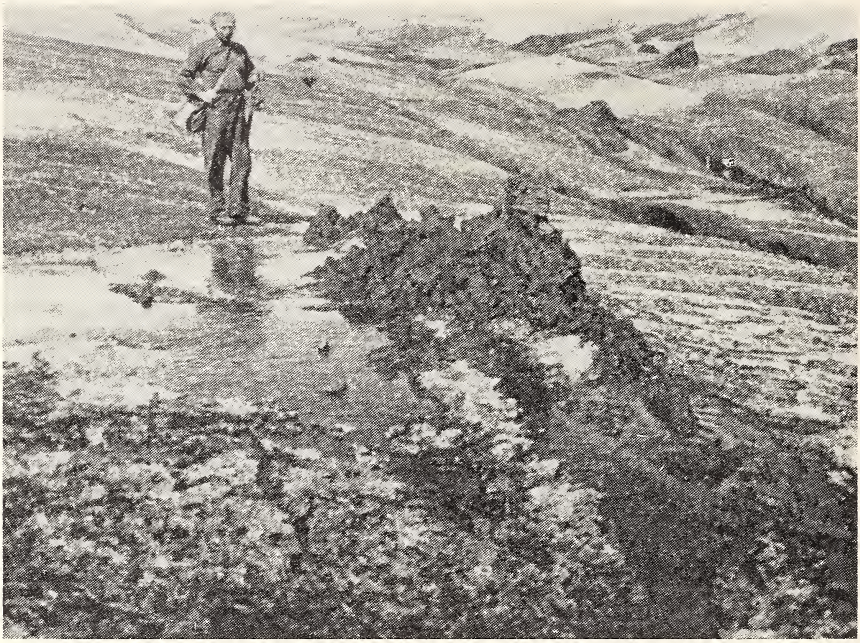

Fig. 13. Transversal moraine on Hoffelsjökull appears as a crest along a narrow crevasse. Its material consists of upthrust basal load.

crevasses (Fig. 13). From such ridges drift spreads over the ice surface, forming accumulations parallel to the glacier margin. Their material is unsorted and to such a degree like that of a basal load that its discussion has been deferred until now.

BASAL LOAD

Fast motion of glaciers and the softness of volcanic rocks, together with irregular relief, enhance glacial erosion and increase the amount of glacial drift and the relative abundance of erosion products in it. According to estimates by Helland (1882), based on the silt content of the rivers starting in the glacier, the area of Vatnajökull erodes annually 0.647 mm. On the same basis, Thorarinsson (1939 b) has carried out a corresponding estimate for Hoffelsjökull, concluding that the floor of the glacier erodes on an average of 5.5 mm a year, or one meter in 180 years. Compared with Jostedalsbrae (Helland 1882) and Aletsch-Gletscher (Collet 1925), the degradation of Hoffelsjökull is notably greater.

Erosion does not, however, take place evenly throughout the glacier floor, for it must be strongest in places where the motion of the basal ice is most rapid and where the floor is most irregular or is composed of less resistant material. From such places the glacier wears away material and drags it along underneath as a basal load as far as the glacier margin or until the friction overcomes the forward-thrusting force. The basal load situated between the glacier and its floor is subject to constant wear, which scratches the stones and causes the stuff to disintegrate. This material, upon being released from the ice, forms basal till (Flint 1947 p. 111).

As the formation of basal load occurs subglacially, it cannot be observed directly. Basal load comes into sight only at the glacier margin and at the basal part of crevasses, and there too it has already become detached from the ice. In other words, basal load originates in the glaciers of Iceland under subsurficial conditions, is transported in the basal part of the ice and mostly also becomes detached already under the ice. Thus the ice melts simultaneously both on the surface and at the base. This is to be observed mostly plainly at the termini of ice tongues, but the same phenomenon recurs also in places situated farther up, as in the mountains of Svinahryggur between the ice tongues of Hoffelsjökull as well as between the mountains exposed along the eastern margin of Hagafellsjökull. In both places the ice covering the melted and detached basal load was clean.

At the southern outlets of Vatnajökull basal load occurs under the snout, at least at Hoffelsjökull, Fláajökull, Heinabergsjökull and Breidamerkurjökull. At all the observation sites the load occured as water-soaked porridge-like till, over which the glacier projects, roof-like. In some places there was a narrow empty space between this roof of ice and the till, indicating that the ice did not press down on the till. The basal part of the ice was there impregnated by the load, for both on the surface of the ice tongue and in the walls of crevasses load could be seen in the ice. It contained glaciated stones, gravel and sand. Among them were rounded cobbles with weak striations (Fig. 14). The load extended in the ice of Hoffelsjökull about 15 m higher than the glacier snout and was limited sharply against the overlying clean ice. The corresponding border on the eastern side of the ice tongue of Breidamerkurjökull was situated at an estimated height of 8—10 m. The upper surface of the ice containing the basal load had a dip of about 30° downward and back toward the glacier, which could be noted from the walls of the radial crevasses. The longitudinal axes of most of the stones stuck in the ice were oriented in the same direction, for from the walls of the radial crevasses it could be observed that the stones lay oriented forward downstream and that the longitudinal axis of most were approximately perpendicular to the margin. In the walls of the crevasses there could also be observed cavities, hollows, that were extended in the same direction (Fig. 5, Plate III). Their flattened and

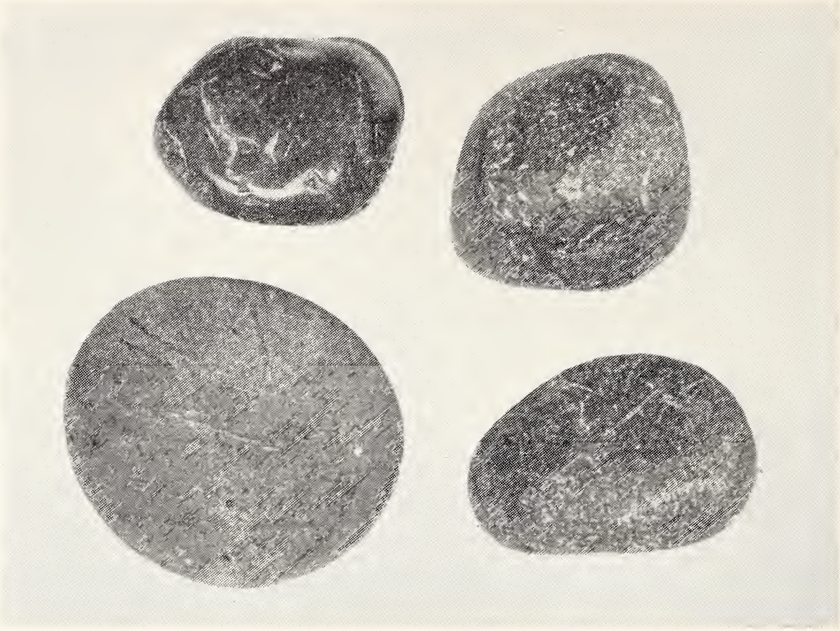

Fig. 14. Weakly striated roundstones picked off the ice on Hoffelsjökull. The size is reduced 1: 2. Photo Erkki Halme.

elongated oval shape indicated that movement continued even at the marginal zone within the ice.

In addition to regular load-impregnation, in the margins of outlets there occur bands of basal load forming transversal moraines (p. 31). They occur both in the overlying and the load-impregnated ice. In numerous places underneath a transversal moraine there is to be found a parallel crevasse sloping upstream full of melted and wet basal load. At the edge of Hoffelsjökull such crevasses were 30—50 cm broad and they were bounded on either side by relatively clean ice (Fig. 6, Plate III). The grain sizes of the load filling the crevasses are shown by the curve 4 (Fig. 6, p. 20). Compared to the grain-size curves for till to be presented subsequently (Fig. 19, p. 42), it proves to be quite identical, revealing the crevasse filling to be basal load in composition. No pollen or spores could be found in the load. The same kind of water-soaked material comes forth also out of transversal bands situated farther from the glacier margin, and sometimes it appears as if the load of a transversal moraine had been squeezed upward out of a crevasse (Fig. 7, plate IV). These crevasses are narrow and in them drift occurs here and there. To judge from the slope of the crevasses, the drift has risen at an angle of about 30° from below up to the surface of the glacier. It is probable that the crevasses bringing forth drift act as shear planes when ice sheets farther back thrust up on top of those in front (Philipp 1920). Still farther from the glacier margin the transversal moraines end. Simultaneously the walls of the radial crevasses turn to clean ice. As certain of these crevasses extended on the ice tongue of Hoffelsjökull open to a depth of about 15 m, or a few meters lower than the level at which the glacier snout was situated, they bore evidence that the rising up of the basal load occurred only at the glacier margin. At the glacier snout there is in places so much drift detached from the ice that it forms a wall-like transversal moraine on top of the snout (Fig. 8, Plate IV). Underneath it the ice melts more slowly than farther back, where there is less drift on the surface.

The rising up of the basal load to the glacier surface takes place on other ice tongues of Vatnajökull as well. During the glacier outburst caused by the eruption of Grimsvötn, the margin of Skeidarárjökull was rent into high walls of ice, in which could be seen numerous layers of englacial load, one on top of the other and inclined upstream. The uppermost of these layers were situated about 40 m above the base of the ice tongue (Nielsen 1937). Similar layers of englacial drift and ice occur also at Breidamerkurjökull, Heinabergsjökull and Fláajökull (Fig. 9, Plate V). The moraine bands thicken there too toward the glacier snout. As this snout often projects eaves-like at an upward slant, with its lower surface accordingly inclined upstream, it is possible that the visible glacier margin actually consists of a sheet of ice situated between load bands. Thus the true margin would be hidden, buried under glacial drift. We shall return to this matter subsequently (p. 85). Be it mentioned here, however, that in many places outside the visible glacier margin there occurs ice covered by glacial drift.

The rising up of the basal load to the surface of the ice in the marginal parts of a glacier cannot be caused by the configuration of the glacier floor, for from the profile presented in Fig. 3 p. 11, as well as from the common shape of the outlets, it is clear that in most cases the glacier floor is either flat or rises toward the glacier snout. At Hoffelsjökull, Breidamerkurjökull (Todtmann 1936 Tafel 18) and Skeidarárjökull (Todtman op. cit. p. 82, Thorarinsson 1939 p. 205) the base of the ice tongue is situated on a level below its snout; but at Langjökull the basal load rises up in bands to the surface of an outlet terminating in a valley sloping outward from the glacier. The basal load rises up in slanting bands also in the ice cliff (Fig. 9 p. 23) of Hagafellsjökull terminating at Hagavatn. The causes for the rising up of the loads are evidently to be found in the movement mechanism of the glaciers of Iceland.

WASHED DRIFT

By washed drift herein is meant glacial drift not only washed by water but also transported and re-deposited on the ice by water.

All glacial drift occurring on the surface of the ice is subject to washing by water. Washing is weak in the accumulation area, where instead of raining it mostly snows and where melting takes place in the main by evaporation. Moreover, the drift accumulating on the surface of the glacier remains visible only a short time. Washing is stronger in the ablation area, which receives more rain water and where meltwater also occurs more abundantly. There is an abundance of water especially at the southern margin of Vatnajökull. To be sure, about half the abundant precipitation takes the form of snow falling during the winter, but since the snow no longer accumulates in the ablation area, the water thereby bound is released annually. Thus of the total precipitation, which e. g. in the budget year 1936—37 was 211 cm (Ahlmann 1939 a p. 52) at the closest observation point of Hólar, 15 m above sea level, the unevaporated water takes part in washing the surface of the glacier. The rain water becomes mixed up with meltwater forming on the glacier surface, especially on sunny days in summer.

In summer the surface of the ablation zone is constantly wet and water flows down the slopes. The water gathers into depressions and forms pools, from which new streams flow down to the next hollow. These little streams bring to mind the streams formed by melting snow in spring. They furrow shallow but steepbanked V-valleys into the surface of the ice. The flowing water does not, however, remain long on the surface but disappears into crevasses, forming little waterfalls. They wear and melt away the ice so as to form round-edged, streamlined, winding pipes leading downward; their course through the blue-green ice can sometimes be followed as far as ten meters, until some roundish projection in the ice blocks the line of vision. With a change in the relief of the glacier surface, the course of a stream likewise changes and the pipe it has made runs dry. Such pipes, of which some were empty, others still in operation, could be seen on the surface of ice tongues in many places. Their diameter was under one meter and nowhere was it observed that a stream filled the whole pipe with water, although the season of thaw was at its height (July—August). Part of the water flows in little surface streams all the way to the glacier margin. By far the greatest part of the water leaves the ice, however, subsurficially.

The water flowing on the glacier surface carries out a vigorous washing of drift for a long time already during its glacier transport. The longer the drift is exposed, the longer it is subject to washing. The heat-binding effect in itself of the dark drift enhances the formation of meltwater even in the part of the glacier where ablation is otherwise slight. This becomes evident upon e. g. studying the aforedescribed medial moraine of the Breidamerkurjökull on the topographical map. On each side of the medial moraine there forms already in the upper part of the ablation zone a narrow valley that collects meltwater, which at every opportunity disappears within the glacier. Into these valleys there also collects drift washed down from the medial moraine, forming sorted, fine-grained sediments on the ice. Mostly they accumulate in the valley on the sunny side, but gradually the washings, mixed up with stones rolling down the moraine, are carried by the water into crevasses and disappear from sight.

Like the medial moraine, the rest of the surface drift is washed. The washing appears to be strongest in the fine-grained material, which is put into motion even by a weak surface flow. The film of dirt formed by eolian drift is washed by heavy rains off the surface of the ice and only after the rain does new dirt begin to spread over the cleansed ice from dirt bands and other eolian material. Eolian drift does not long remain, therefore, in the place where it has accumulated or where it becomes exposed out of the ice in the ablation area; but on the surface it becomes subject to both washing and transportation. Likewise, the water also treats the ash layers and transversal moraines on the surface, but since the stuff of which they are composed is coarser than in eolian accumulations, the water detaches from them only the finer part, while the coarser material is transported only a short distance or stays in place.

The material loosened by the surface washing travels with the water into pools on the glacier surface, where it forms sediments. Thus behind the marginal ridge at Hoffelsjökull a small delta (Fig. 10, Plate V) had formed on the glacier surface out of drift sliding down from the transversal moraine. Its proximal side faced the marginal ridge, i. e. in the distal direction of the glacier, and its layers sloped toward the saddle situated behind the marginal ridge. The material of the delta was clearly stratified. Its sorting and grain size are shown by the curve 5 in Fig. 6 p. 20.

The material was in the main fine sand, but mixed up with the fine material there occurred here and there scattered stones that had rolled down the border crest. In addition to the mineral matter the sediment also contained 1.26 % organic matter, by weight. This figure included pollen and spores from the following plants: *Betula* 4, *Cyperaceae* 1, *Graminae* 1 and *Selaginella* 1.

A water sample taken from the small pool formed on the surface of Skaftafellsjökull, measuring 6 × 4 m across and 30—40 cm in depth, contained an abundance of *Tabellaria flocculosa* and other small diatoms of shallow, fresh water (Mölder 1951, p. 132), which accordingly in this case grew in the pool on top of the ice. There had sunk to the bottom of it sediment consisting of sand and organic matter, including e.g. brown leaves of *Betula nana* and *Salix*.

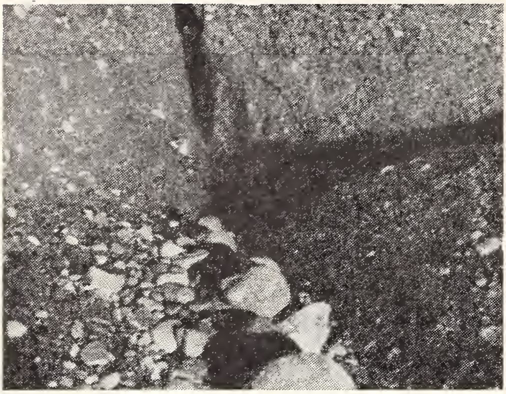

Fig. 15. A row of washed stones at the bottom of a furrow cut into the surface of Breidamerkurjökull by a superglacial river. The stones are 5—10 cm in diameter.

The water in streams starting in pools is always clear but it collects from the glacier surface new drift, which accumulates in the still waters farther on. Thanks to the sedimentation basins the water of the surface streams remains clear even when it runs into a crevasse or flows over the glacier margin. At times stones roll down off the ice into the valleys worn by streams; and these stones are washed and form a regular line at the bottom of the furrow (Fig. 15). As the surface relief of the glacier changes, the streams seek new channels and new sedimentation basins. The old V-valleys and pools run dry and their sediments begin to retard the melting of the underlying ice. The sediments sinking originally to the bottom of depressions are thereby eventually left at a level above the surrounding surface and in turn begin to wear away. Thereby accumulation and erosion alternate at the very surface of the ablation area, and the surface drift undergoes ever new sorting. Its accumulations begin to contain in ever-growing measure material of different origin, which stage by stage travels toward crevasses and the glacier margin.

The occurrence of washed drift on the glaciers of Iceland is not limited solely to the surface, for sorted material also appears from within the ice. On the vertical wall of Hoffelsjökull's snout it formed a lens, which appeared to continue into the ice. (Fig. 16 a). The material of the lens was sorted to the extent that its basal part was composed of gravel with pebbles, which

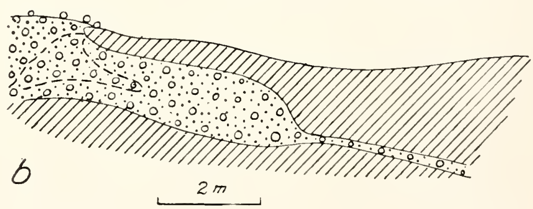

Fig. 16. Two lenses of stratified and sorted drift in the ice: a. in Hoffelsjökull, b. in Heinabergsjökull. The melted part is separated by a dashed line from the frozen material.

farther up changed to gravelly sand. Its grain sizes and sorting are indicated by the curve 6 in Fig. 6 p. 20.

The lens was covered by a thin layer of ice, which contained a weak impregnation of drift. At one point ice was lacking and there was sand instead, which in turn was covered by stratified sediment of fine sand. Overlying it like a roof rested a flat-bottomed stone.

In the marginal ice of Heinabergsjökull there occurred a sand and gravel lens about 4 m long, the end facing the glacier of which formed a narrow, downward sloping sheet (Fig. 16 b). The end facing away from the glacier was rounder and reached the surface. The core and rounded end of the lens had melted, but otherwise the sand and pebbly gravel within remained in a frozen state. No flowing water could be seen in this lens either, which signified that the material had to be sorted either during the stage of glacier transport or before its becoming embedded in the ice.

Washed drift works loose on the glaciers of Iceland for the most part through the action of subsurficial meltwaters. Extensive outwash plains, or sandurs, have been formed out of this material outside the glaciers. In some places, as at the margin of Heinabergsjökull and Breidamerkurjökull, the outwash plain has partly accumulated on the ice, for as the rivers flowing outside the visible margin of these glaciers wear away the outwash plain, a firm sheet of ice is exposed below (Fig. 11, Plate VI). This ice is apparently stagnant and melts in place, for which reason the drift layers formed on top of it can no longer be regarded as participating in the glacier movement. On the other hand, these layers are not wholly independent of the glacier action either, for they become deformed upon the ice's melting underneath them; and the accumulation built of them attains its ultimate forms and construction only after the wasting away of the ice. The accumulations covering the glacier margin thus constitute an intermediate stage between the glacial drift participating in the glacier movement and that accumulated outside the glacier.

CREVASSE FILLINGS

The crevasses of the Icelandic glaciers tend to be divided, according to the locality of their occurrence, into three groups: crevasses occurring in the upper parts of ice tongues, those occurring in the outermost parts and those in various parts of the area in between.

The crevasses in the upper part of ice tongues are long and run perpendicular to the slope of the glacier (Fig. 17). These transversal crevasses move with the ice downstream, but they do not last long, for they close up from the effect the glacier flow, the freezing of the water and snow collecting in the crevasses.

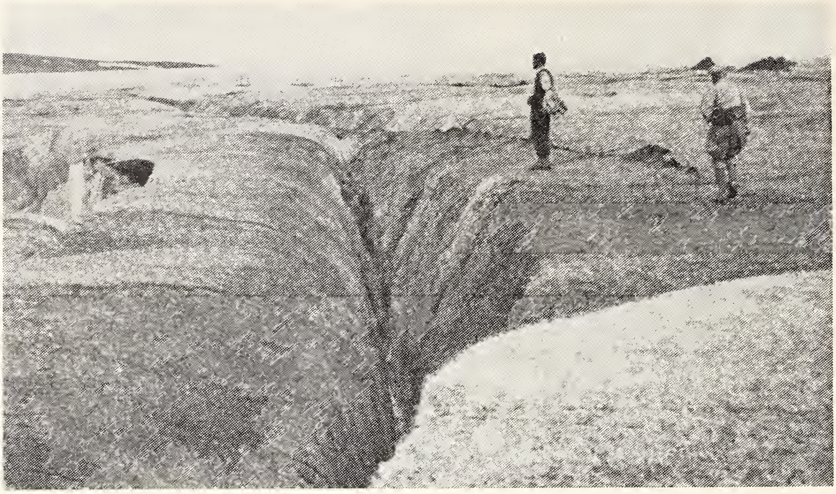

Fig. 17. Transversal crevasses crossing East Hagafellsjökull. The mountain Hagafell in the background. Photo Helgi Jónasson, 1946.

Localized crevassing occurs superglacially on ice tongues farther downstream also at points where some part of the glacier advances faster than the ice around it. On the surface of the glacier there open up at right angles to the flow crevasses which close upon moving downstream. At such places the surface of the glacier tends to break up into large blockfields of ice. As these points remain stationary on the ice tongue, the cause of the phenomenon must be the topography underlying the glacier.

Two types of crevasses occur at the terminus of an outlet. One type is oriented, as described previously (page 34), forming narrow transversal cracks. The other type consists of open radial crevasses (Fig. 18), which get narrower upstream and terminate a few score or at most a few hundred meters from the ice border.

The crevasses of every category collect water except the transversal ones filled with drift. The drift is also washed into the crevasses from the ice surface by water. Only the drift collecting the radial crevasses travels without obstruction down to the base of the

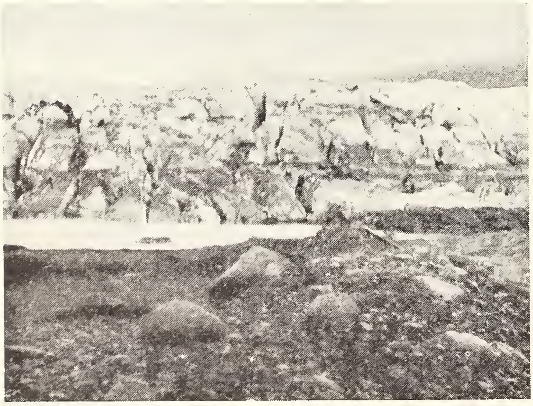

Fig. 18. Radial crevasses in the terminus of Fláajökull. In front a radial moraine separated from the glacier by a lateral stream.

glacier, for these crevasses often extend down through the entire ice sheet. At the base of such 10—15 m deep open crevasses at the snout of Hoffelsjökull there was silty, stagnant or slowly flowing water, into which stones kept falling off the ice and from the crevasse walls. The water had melted the lower parts of the crevasses into cavelike extensions, the level of the water filling the bottom of which reached approximately to the same level as the marginal lake in front of the snout. Inasmuch as the water level in at least a couple of crevasses that did not extend to the glacier margin was the same as in open ones, the crevasses may have been connected subglacially. No sample could be obtained of the sediment deposited in the water, but at those points where the base of the crevasse reached above the waterline, it had an even surface (no other kind of relief could be preserved in such wet material) and its filling consisted of a »mushy» mixture of sand, stones and in places also till. The same kind of sediment occurred likewise in the open radial crevasses of Breidamerkurjökull, at least at those points where the base of the crevasse appeared above water.

The open radial crevasses occurring at the ice margin of Fláajökull deviated from those described in the foregoing in that the free water was lacking at their base. These crevasses were more than 10 meters deep and many of them reached to the glacier margin. One of these crevasses was 1—3 meter broad at the bottom and the base of its walls consisted of ice containing basal load at a height of about 1.5 meters, measured from the bottom of the crevasse. At the base of the crevasse there was a layer of basal till at least 50 cm thick, composed of silt and fine sand (Fig. 12, Plate VI). There occurred no water-washed drift whatsoever at the bottom of the crevasse. The basal till at the bottom of the crevasse was so mushy that the stones on its surface gave way underfoot, sinking into the till. The till appeared to be squeezed into the crevasse from underneath the bordering ice walls, for these ice walls rested on top of the thawed-out and mushy till, pressing it down. The boundary between the ice and the underlying till was sharp and approximately horizontal, but was curved upward at points where boulders had become detached. The ice was thus melting from below here too and discharging its basal load. At the bottom of the crevasse the till reached 20—50 cm higher than the boundary between ice and till on either side.

Behind the terminus of the outlet the ice is already so thick (at Breidamerkurjökull as much as 470 meters), that the crevasses opening up at the surface can no longer extend through the ice (comp. Flint 1947 p. 17). This conception is supported by the fact that the crevasses do not stay open but close up on moving away from the fracturing area. The surface fractures extended to a depth of at least 10 meters, clean-walled without any accumulations of drift in view. Evidently the surface drift washed down into the crevasses by water works down even deeper, disappearing from sight. Simply by inspecting the crevasses it is impossible to decide what becomes of this drift when the crevasses close up again. If it remains in the crevasses the drift forms sorted lenses and strata within the ice. These formations travel invisibly with the ice and are exposed only after the effects of ablation. The occurrence of glacier breccia below ice falls and the exposure lenses formed out of sorted material at the glacier margin points to such a process. If, again, drift is washed by water seeking a downstream course through the plastic part of the glacier and the basal load-filled bottom layer, it changes character: from cervasse filling it first alters into washings, transported through tunnels, and then settles either inside the tunnels or outside.

MORAINE DEPOSITS

In glacial research concerned with the Pleistocene epoch moraines have commanded attention from the very start. The term designating these formations derives from a French word used by the inhabitants of the Western Alps for deposits of glacial drift in front of valley glaciers, while its meaning is given various nuances. In German, Scandinavian and Finnish literature the word *moraine* is used to designate both the formation and the material composing it. British researchers have limited moraine to signify, in particular, the morphological accumulation, while the separate term »*boulder clay*» is used for the material. American researchers observe

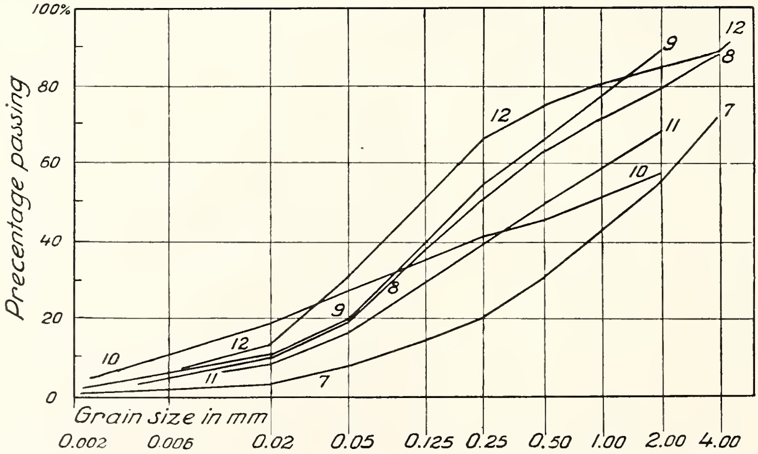

Fig. 19. Grain size distribution in the glacial drift of moraines. Curve 7 ground moraine at Breidamerkurjökull, No. 8 the same at Langjökull, No. 9 d:o at Hagavatn, No. 10 push moraine at Svinafellsjökull (Hoffelsj.), No. 11 end moraine at the river Hanypá, and No. 12 the same at Hoffelsjökull.

a similar usage, but they have substituted for the English boulder clay the Scottish word »*till*». According to Flint (1947 p. 126), »moraine is accumulation of drift an initial topographic expression of its own, built within a glaciated region chiefly by the direct action (deposition and thrust deformation) of glacier ice».

Moraines are conceived in this study, too, in accordance with Flint's terminology, largely as morphological accumulations of glacial drift forming as a result of glacier action both on the ice and beyond it. Among the former are included medial, lateral, and tranversal moraines, dealth with earlier, which take part in the glacier motion and are usually deformed when ice melts underneath them. Inactive moraines are grouped in the following as ground, radial, ablation and end moraines, each of which type has its own morphologically distinct features. Accordingly, the material composing the moraines has not determined the classification, but it will be dealt with in connection with each moraine type.

GROUND MORAINE

By ground moraine is meant in the following glacial drift directly freed from the ice, forming a regular accumulation or one conforming to the topography of its base. Such moraines cover broad areas in the highland of Iceland, but in front of ice tongues they are encountered only in small areas recently exposed from under masses of ice. In such places the ground moraine sometimes extends up to the ice border. A small accumulation of this kind occurs e. g. in front of the northeastern border of the snout of Breidamerkurjökull south of the Fellsfjall mountains. The surface of the ground moraine is almost flat but here and there occur small boulders and stones (Fig. 13, Plate VII). A large part of these are angular and striated glaciated stones, in addition to which scratched roundstones are to be found. In between the stones and boulders the surface consists of unsorted both coarse and fine sandy till, which has dried on top but become saturated with water already at a depth of 50 cm. Toward the ice margin the ground moraine became moist also on top and continued as a thick, porrige-like mass under the terminus of the glacier in such a way that the snout overlies this mass without pressing it down. The undersurface of the ice was, as could be observed from the marginal crevasses, approximately horizontal and was situated on the same level as the surface of the ground moraine in front of it. Since the ice overlying the ground moraine contained only a little glacial drift, the ground moraine had here been formed out of the basal load of the glacier. It had become detached from the ice already subglacially and remained in place undisturbed while the ice melted above it.

A grain-size analysis (Fig. 19) made from a sample taken from the ground moraine at a depth of 20 cm shows the ground moraine to consist of unsorted till. The main components are coarse and fine sand, but other grain sizes occur as well. The fairly even distribution of fractions and the presence of fine grains prove the ground moraine to be unwashed.

At the point where the sample was taken, a till fabric analysis was also carried out (Fig. 20 a). It shows the long axis of the stones to have a distinct maximum, pointing toward the nearby glacier. The orientation of the stones in the ground moraine thus shows in this case the direction of glacier flow; and because the ground moraine is in an undisturbed position, it is obvious that its orientation dates back to subglacial conditions under which the basal till still moved along with the ice.

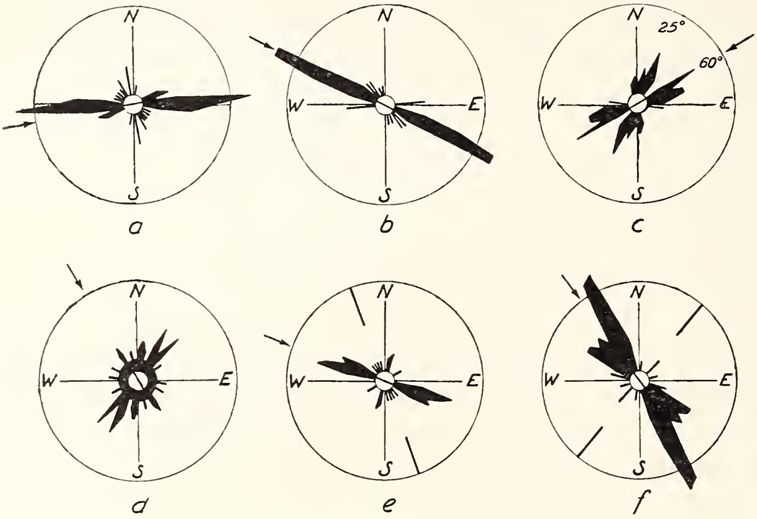

Fig. 20. The orientation of stones in moraines: a. ground moraine at Breidamerkurjökull, b. the same at Heinabergsjökull, c. d:o at Langjökull, d. end moraine at Fláajökull, e. end moraine at Heinabergsjökull, f. end moraine at Breidamerkurjökull. The radius of the circle indicates an orientation of 20 %. Arrows signify ice-movement direction.

The freeing of ground moraine from the ice may be observed also at the termini of Heinabergsjökull, Fláajökull and both tongues of Hoffelsjökull, where a ground moraine landscape is developing between the ice and frontal moraine. The basal till becomes exposed from underneath the glacier as a very wet mass detached from the ice by melting; it stays in place only in the event that it dries. On a sloping base it tends to form solifluction earth. Ground moraine dries still less readily in the trough between the frontal moraine and the glacier where meltwater often collects. The ground moraine there forms a water-soaked mass, which in addition to till contains sorted sediments and superglacial drift. If the glacier margin projects eaves-like over the drift, large fragments of ice break off, causing upon falling down the formation of ridges around it (Fig. 14, Plate VII). The imprints left by the ice fragments do not last long, for the mushy mass flows into the hollow produced by the melting of the ice and fills it, whereupon the flat surface topography characteristic of ground moraine is restored. The grain components and structure of the »till dirt» as well as the orientation of its stones naturally deviate from the features of ordinary ground moraine, even though the accumulation morphologically corresponds most closely to ground moraine.

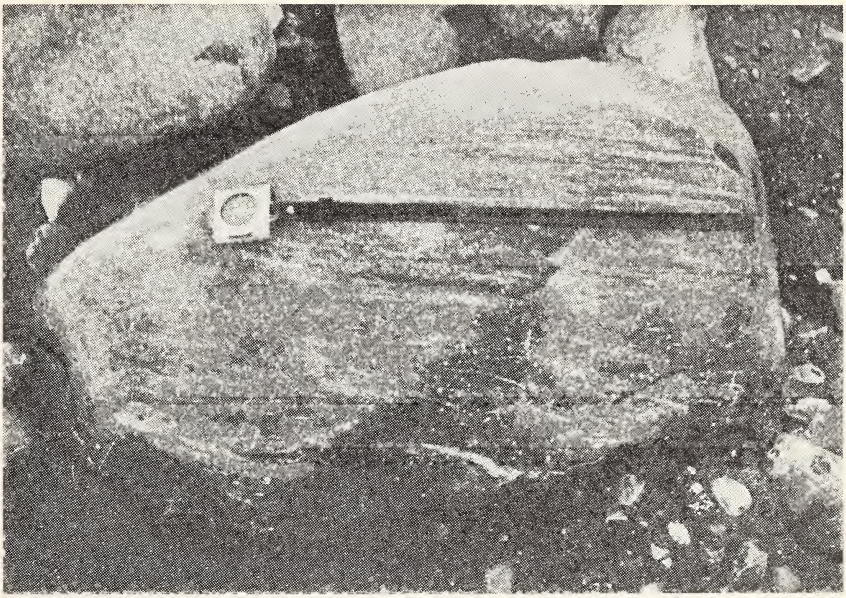

Fig. 21. Striated boulder from the ground moraine surface in front of Heinabergsjökull. Ice has moved over it from left and shaped stoss and lee sides onto the boulder just like outcropping rocks. Boulder is 70 cm in length.

The ground moraine in front of Heinabergsjökull is exposed to view at the bottom of a dried river bed. It is composed of till with an abundance of stones and boulders. They are fixed within a clayey and densely packed mass. The boulders have been polished by ice and constitute small »roches moutonnées» with stoss and lee topography and with overall striations (Fig. 21). These »roches moutonnées» have a very distinct orientation, for their stoss sides face upstream and their lee sides in the opposite direction. This can hardly be explained otherwise than that the glacier had passed over the ground moraine without destroying it and worn boulders projecting up out of the surface into »roches moutonnées». Their orientation analysis

(Fig. 20 b p. 44.) shows the longitudial axes of the blocks to be in the flow direction just about as distinctly as in the basal till in front of Breidamerkurjökull.

The structure of ground moraine becomes evident in the walls cut by the Skaftafellsá river in front of Skaftafellsjökull. The material of the ground moraine is dense and stony basal till, and in the approximately 3 meters high wall there could be seen no kind of stratification whatsoever. The ground moraine extended at least to the waterline. The thickness of the layer could not be determined.

At the margin of Fláajökull the ground moraine consists of basal till, which extends under the ice. The terminus of the glacier here clearly presses the underlying basal till, which has melted free of the ice, and squeezes it on a slant underneath the margin. The boundary between the basal till and the overlying ice is sharp and dips downward toward the glacier. At the margin one gets the impression that the terminus of the outlet rises upward and that the base of the ice sheet beyond the outlet lies below the level of the snout. From underneath the glacier a ground moraine landscape sloping toward the glacier is thus becoming exposed inside the arcuate end moraine.

In the central highland where the glacier is free from surface drift almost as far as its terminus, the ground moraine consists almost exclusively of basal till. Angular and striated glaciated stones occur both on top and inside the ground moraine. The fine stuff composing the till is clearly unwashed and unsorted. Furthermore, the stones are orientated clearly parallel to the glacier movement. Thus it does not deviate in structure from the ground moraines occurring elsewhere. Attention is arrested, however, by the fact that the ground moraine forms an unexpectedly thin layer on the surfaces recently freed from ice. Accordingly, on the eroded and polished rocks of Svinarhyggur exposed by melting out of Hoffelsjökull in recent years there occurs glacial drift only in depressions in the rocks and spaces between knobs, forming beds at most 50 cm thick; and higher surfaces there are only little stones and blocks (Fig. 24, Plate XII). Since the surface has been washed only by rain water since its exposure, the scantiness of the ground moraine must be caused by the fact that even originally the ice had not brought more than a thin layer of basal till on to the rocks. On the other hand, the ice had given them a stoss and lee topography as well as glacial striae.

The ground moraine covers the smoothed rocks outside the glacier margin at the eastern end of Hagavatn just as thinly; and the surfaces of the rocks have erosion marks as well as the products of mechanical weathering. At numerous points stones and blocks were scattered atop the rocks, lying as if atop the erosion marks they had caused. Only in sheltered places did there occur a uniform moraine, at most a meter thick. On the slope descending toward the wilderness camp of Hagavatn, this ground moraine contained an abundance of fine sand and silt and was lacking in blocks. On its surface could be seen stripes consisting of small stones, indicating solifluction.

In the valleys and dells of the central plateau the ground moraine is thicker. It forms a layer estimated to be 1—1.5 m thick in the upper part of the valley which separates the group of aforementioned mountains from the basin of Hagavatn. This valley was freed of ice only as late as very recent decades, for there was ice there in 1929 (Wright 1935 Plate I) and even as late as 1949 the margin of Langjökull extended into the northeastern part of the valley. The surface of the outlet slopes gently toward the terminus and buries itself in the ground moraine, with the result that the true position of the margin remains unknown. The ice cannot extend very far underneath the drift, however, for about 300 meters from the visible border of the glacier polished rock surfaces crop out of the ground moraine, which »climbs» up their gentle slopes. Otherwise the surface of the ground moraine is almost flat and with a gentle slope parallel to the length of the valley. Around the rocks the ground moraine is looser and drier on the surface than the parts deeper down. The grain composition shows it to be unsorted basal till (Fig. 19 p. 42). An orientation analysis carried out at the same spot (Fig. 20, p. 44) gave the longitudinal axis of the stones two maximae, the compass reading of one being 25° and of the other 60°, though the longitudinal axes of many of the stones deviate from these directions. Of the maximae one (60°) was oriented approximately parallel to the length of the valley toward the end of the outlet and evidently corresponded to the last course of the glacier flow, whereas the other (25°) pointed straight toward the glacier proper, hidden behind the mountain earlier exposed out of the ice.

The rock outcrop which the present writer examined only after performing the orientation analysis, was eroded in such a way that the stoss side, with its prominent striations indicated that the glacier had advanced over the rock from a 30° direction, thereby eroding the rock heavily. Weaker striae occurred furthermore with a 60° orientation, crossing the system of the main direction. On the basis of these cross-striations it appears certain that the ice moved over the area first from the direction of 30° and then of 60° and that the earlier stage of advance eroded the rock more than the later, which was unable to destroy the erosion marks of the earlier advance from the relatively soft rock surface.

The striae on the rock surface thus have almost the same compass directions as the orientation maximae of the till. It is accordingly evident that the advances of the glacier are reflected by the compass readings yielded by the till. The basal till thus had to be formed already during the main glacier advance and the last oscillation of the glacier had only deformed the earlier ground moraine by turning the stones in the direction of its own motion.

In order that the glacier could have advanced over the observation site from the 30° direction, it would have had to be thick enough to bury the mountain now separating the observation site from the present glacier proper, or considerably thicker than the last ice tongue to survive in the valley. Observation thus shows the strongest erosion and basal load transport to have taken place underneath relatively thick ice. The erosion marks and till orientation created at that time remain even though the ice had subsequently advanced over the same place from another direction. Corresponding observations made in Iceland have been previously presented. Thus Kjartansson (1939) observed numerous crossing striae in southern Iceland, which are grouped into two systems of different age in such a way that the erosion forms reflecting the older, general and extensive glaciation are more distinct than the younger striations caused by later, local glacierizations.

It is natural that during the stage when the glacier erodes the rock most it also carries off the most load from its floor. The transport of basal load is thus associated genetically with the activeness of the glacier. Load in greatest abundance forms in the place where the glacier erodes its floor most. From this it follows that chronologically the greatest abundance of basal load is created during the stages of glacier advance.

The release of basal load from the ice takes place subglacially and in many cases accumulations of it remain in place undisturbed, preserving their original structure. At the ice margin it continues directly in the form of ground moraine, which mainly constitutes an accumulation of basal till formed out of basal load.

RADIAL MORAINES

In front of Breidamerkurjökull, Hoffelsjökull, Heinabergsjökull, Fláajökull and Svinafellsjökull radial moraine ridges, which often represent extensions of crevasse fillings described earlier, occur on the surface of the ground moraines. In front of Breidamerkurjökull they are evidently rapidly vanishing miniature forms, but the radial moraines in front of Heinabergsjökull, Fláajökull and Svinafellsjökull are so large and firmly built that they represent permanent accumulations.

On the ground moraine accumulated along the eastern border of the outlet of Breidamerkurjökull the radial moraines form sharp-crested ridges continuing outward from the glacier and gradually becoming lower (Fig. 15, Plate VIII). In length they are at most 10 meters and in height 0.5 meters, and at their base they measure about a meter in breadth. The main orientation of the ridges is at right angles to the glacier margin, but often the ridges run together and sometimes there are transversal ridges between them. Between the ridges there are angular ground moraine hollows. The ridges are best developed in the part along the immediate ice edge, and they become lower and the slope becomes more gradual as the distance from the margin increases.

The material of the radial moraines consists of the same kind of basal till as is contained in the surrounding ground moraine. There is the difference, however, that the material of the radial moraines is unoriented.

The glacier margin against the area of radial moraines is fractured by open longitudinal crevasses to such an extent that it breaks up into blocks of ice (p. 45). Radial moraines are in this case produced in the spaces between blocks of ice and the hollows between ridges represent the imprints of fallen ice blocks. The radial moraine levels out and disappeares from the surface of the ground moraine unless both dry rapidly. In the event that the ice contains glacial drift exposed on the surface by melting, this also forms part of the material of the radial moraines, because the surface drift slides and is washed down into the crevasses.

In the terrain in front of Heinabergsjökull the radial moraines deviate from those described in the foregoing. They form broad, flat-surfaced ridges between the old end moraine and the ice margin and continue as coherent but uneven zones for several hundred meters. A zone of this kind stands out from its surroundings not only on that it has an elevated surface but also on account of the abundance of boulders occurring in it. These boulders measure 1—2 meters in diameter, are angular and situated loosely on the surface. The material between them is gravelly and sandy till (Fig. 10, p. 28). The surface layer of till is loose to a depth of about 0.5 meters and often there is a thin layer of coarse sand under the stones. There appears to be no regularity in the lengthwise orientation of the blocks, for their longitudinal axes were turned every which way. On the other hand, the fine material indicated a weak maximum in the compass direction of 245°, which at the same time was the longitudinal orientation of the ridge. The glacier lay in this direction, too, looked at from the observation site. The ridge does not, however, extend as far as the glacier, but it is cut off by a lateral stream. On the other side of the stream there is irregular terrain, as an extension of the ridge, on the surface of which here and there large boulders occur. Such boulders were to be met with even on the surface of the ice, for just here a medial moraine appears on the surface of Heinabergsjökull. The surface of the glacier has remained elevated under the moraine and in addition it has suffered considerable fracturing. It is natural that this radial moraine has been built up out of the surface drift washed from the medial moraine.

In the terrain in front of Fláajökull radial moraines occur in abundance (Fig. 22). They start from the old terminal moraine and gradually descend the gentle slope as parallel ridges toward the glacier until here too a lateral stream breaks their descent.

Because of its winding course the bed of the stream sometimes runs along the very margin of the ice and sometimes it runs so far from it that a strip of fluvial deposits is left between stream and glacier. In places where the stream runs along the ice margin, the radial moraines extend nearest the edge of the glacier broken up by longitudinal crevasses. The fillings of these crevasses have already been discussed (p. 41). The radial moraines measure 10—200 meters in length, 0.5—1 meter in height and about 2 meters in width at the crest (Fig. 18 p. 40). The stony and block-filled material of these radial moraines differ from that of the radial moraines of Heinabergsjökull in that here the stones have been scratched by the ice, though, to be sure, some of them are rounded. Stones occur in greatest abundance on the surface, for already at a depth of 5—10 cm the material turns to dense, silty and clayey basal till, with stones here and there. It continues in this composition at least to a depth of 70 cm. A fabric analysis of the ridge carried out at a depth of 50 cm shows a weak maximum in the directions of 210° and 225°, but the distribution in all directions of the compass was so great that the material may be regarded as lacking in orientation. Inasmuch as the orientation of the ridge (as well as the apparent

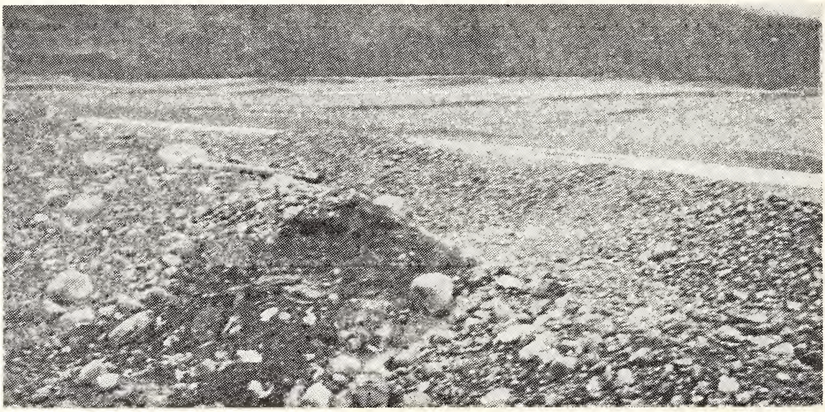

Fig. 22. Radial moraines in front of Fláajökull, seen from the end moraine. The glacier terminates about 200 m west (left).

final direction of advance of the glacier) is 335°, the maximum orientation of the stones lies obliquely in relation to it.

The identical composition of the material of the crevasse fillings and the radial moraines indicates an identical origin, *i. e.* the radial moraines are in this case crevasse fillings squeezed out of basal till into radial crevasses. This conclusion is supported by the fact that the crevasses and the radial moraines run parallel and that the distance between the crevasses and between the moraines is approximately equal. As the crevasse fillings have formed on a sloping surface, they have dried and survived in the exposed area.

A fourth group of radial moraines occurs in the terrain in front of Svinafellsjökull, the western outlet of Hoffelsjökull. In part, they form islands in the lake created in front of the glacier; in part, they run along the shores of the lake in the form of wall-like ridges; and, in part, they border dry pools with flat bottoms filled with a sediment of sand and silt. The form and orientation of the ridges varies considerably, but in general they are oriented at right angles to the glacier margin, and they consist of series of short, sharp-crested mounds. In form these rows of ridges often resemble small eskers (Fig. 16, Plate VIII). The mounds are 10—15 meters long, 5—10 meters broad, and their peaked summits rise up 1—4 meters above the surrounding sediment.

The surface of the mounds consists of fairly evengrained gravel and small stones as well as an abundance of pebbles (Fig. 17, Plate IX). A close examination of the stones reveals that many of them have been scratched by ice, whether they be angular glaciated stones or roundstones. Inasmuch as the material of the ridges changes immediately below the surface so as to contain fine-grained fractions as well, it must have been freed from the ice directly. It accordingly corresponds to the drift occurring in the ice of the eastern outlet of Hoffelsjökull, which in addition to glaciated stones contains roundstones. The glacier evidently advanced over glacifluvial deposits and carried away load with it. If, on the other hand, the drift of lateral moraines had been transported by streams of meltwater only after becoming liberated from the ice, all of its stones would have had to erode and become rounded, in which case the striae possibly occurring on them previously would have disappeared.

The ridge area of Svinafellsjökull is situated on the bottom of the lake once dammed up between the glacier and the end moraines in front of it but subsequently drained. The conditions prevailing during the origin of the topography had therefore been subaquatic, although the ice could not float in the shallow water. In extending into the water and disappearing therefrom, the ice tongue had apparently been broken up by radial crevasses. The radial moraines formed in these partly water-filled crevasses. Both the water-filled and dry silt and fine sand-bottomed pits between the moraines indicate the spots where the ice wasted away last. The loose structure of the lateral moraines indicates, moreover, that their material consists of till from ablation moraines caving in down the crevasses. Upon the melting of the ice bordering the crevasses, the crevasse fillings collapsed, causing the formation of steep slopes and sharp crests. The whole accumulation might thus be regarded as belonging to the ablation moraine; but, on the other hand, the orientation of the ridges together with a morphology identical with the radial moraines of Breidamerkurjökull and Fláajökull, described in the foregoing, speaks in favor of their being classified as radial moraines.

The observations reported in the foregoing represent all the radial moraine types encountered by the author in Iceland as regards the conditions of origin of which it was possible to drawn a definite conclusion. Accordingly, radial moraines form in Iceland during the melting stage from both medial and lateral moraines as well as crevasse fillings. In the former instances, they produce a broad, irregular zone, whose material is rich in blocks and unwashed. In the latter instance, the radial moraines form in the radial crevasses of the glacier margin in two different ways: 1- surface drift avalanches into the crevasses, and 2- mushy basal till is squeezed up into them. Furthermore, blocks of ice breaking off the snout of glacier outlets and falling into the mushy ground moraine throw the till up to form ridges, which under suitable conditions remain.

Radial moraines thus form on the very glacier margin, where the material embedded in the ice is released by melting. Inasmuch as the glacier motion determines the location and orientation of both medial moraines and radial crevasses, the orientation arising during this motion becomes evident also in the direction in which radial moraines run, even though the ice might have turned stagnant at the time of their formation.

ABLATION MORAINES

In the marginal parts of the glaciers of Iceland there occurs, as pointed out previously, an erichment of glacial drift on the surface of the ice. This is caused partly by the fact that the material coming onto the glacier remains on the surface of the ice and partly that drift embedded in the ice becomes exposed. Rapid melting of the surface belongs among the characteristic features of the margins of Icelandic glaciers. This is because the ice tongues project far below the climatic snowline. Because the glacier motion slows down upon the ice mass's getting thinner, there occurs stagnant ice at the termini and flanks of outlets. The ice disappears most slowly at places where the surface drift forms such thick layers that they retard the melting of the underlying ice. Upon the ice's gradually melting beneath the drift, the accumulations of drift are deformed and produce irregular ablation moraines. According to Flint (1947 p. 131), »ablation moraine is a deposit of drift let down irregularly from the surface of the glacier onto the ground by gradual ablation of the intervening ice». If surface drift does not form, ground moraine appears from underneath the ice; hence, ablation moraine does not occur everywhere in the area of so-called dead ice.

The formation of ablation moraine is to be observed in many places in Iceland. Especially at the northern end of Vatnajökull it covers an extensive area in front of Bruárjökull (Woldstedt 1939, Todtmann 1951, 1952, Hoppe 1953). Extensive ablation moraines have formed also in the vicinity of Langjökull during recent decades (Nørvang 1937). Likewise, they have made their appearance in northwestern Iceland, where the local glaciers have melted away altogether (Keilhack 1934).

Ablation moraines form also at the southern margin of Vatnajökull at places where there is an abundance of surface drift on the ice. Their formation is further to be observed at the margins of Langjökull. As the ice melts under overlying drift only very slowly, it is not possible to ascertain in a single summer how the entire development takes place in one area, but its study requires prolonged work on the same site. Some sort of conception of the origin of ablation moraines can, however, be obtained by quickly examining the gradual transformation of the lateral moraine at the side of an ice tongue into an ablation moraine. Such a developmental sequence could be observed e. g. along the eastern margin of Hagafellsjökull and the western margin of Skaftafellsjökull.

Along the eastern margin of Hagafellsjökull, lateral moraine appears at an elevation of 800 meters above sealevel. Here it contains alternate layers of snow and drift. By degrees the snow disappears from the border zone, now composed of a uniform layer of drift. Only in places where the stream flowing in the zone between the mountain and the ice erodes openings into the lateral moraine is it possible to see that underneath the drift there is hard ice. The surface of this ice was situated higher than the surface of the outlet in front of the lateral moraine and between them occurred crevasses. The ice underneath the drift had thus evidently become detached from the glacier proper and was melting *in situ*. Near the lake of Hagavatn the snow disappeared between the ice and the surface drift. At the same time the lateral moraine became irregular on its surface, where there began to appear kettle holes with steep sides and which measured less than 10 meters in diameter and 2—4 meters in depth. The walls of these holes consisted at the top of drift and below of hard ice. Often there was water at the bottom, which melted the ice at the same time as drift slid down the sides into the hole. Gradually the holes widened and became connected together, whereupon there remained between them sharp-crested ridges of ice and overlying drift. Drift soaked in meltwater flowed slowly down the slopes to the bottoms of the holes, bringing the new, underlying ice into view. The irregular topography had thus been produced by the ice underlying the drift. The drift mantle at the same time underwent perforation. As a consequence of the exposure of the ice, its melting became accelerated, with a concomitant increase in the volume of meltwater. Gradually the regular drift surfaces disappeared totally between the kettle holes and the formation changed so as to consist altogether of interconnected broad hollows and the ice ridges in between. Somewhat farther from the glacier the ice ridges vanished from between the holes, and in place of them there developed the deepest holes of all. The relief thus became contrary to what it had been some distance away. In other words, the bottoms of the holes first formed remained elevated on account of the deposited drift in relation to the new hollows resulting from the melting of the ice last to be exposed. At the same time the topography became more even, because the difference in elevation was slighter and the slopes gentler than in the area containing ice. As Hagavatn's ablation moraine is situated on a slant base between the mountain side and the bottom of the valley, it has been subject to erosion, for its mushy moraine flows down the slope, in addition to which the little brooks streaming down the mountain also act erosively.

A corresponding developmental sequence was to be observed earlier on the shore of Hagavatn south of its present outlet, but in 1949 only an ablation moraine free of ice any longer remained. According to Helgi Jónasson, a native Icelander, there had been a relatively flat-surfaced lateral moraine in the same spot 10 years ago, which at first had become irregular, with ice becoming exposed at the lower ends of the sides of the resulting holes (Fig. 18, Plate IX). The margin of Hagafellsjökull had already retreated from the lake basin, so that the drift-covered ice had become detached from the glacier proper already earlier. After the ice had melted underneath the drift, the topography gradually levelled out.

The lateral moraine at the western margin of Skaftafellsjökull is also turning into an ablation moraine. The lateral moraine is situated at first on top of the ice between the living ice tongue and the bordering mountain, but gradually it runs off the margin of the glacier to follow the foot of the mountain in the form of a fairly flat-surfaced drift zone. Underlying the drift here too is a layer of ice which begins to melt only at the side of the visible terminus of the ice tongue. The melting is promoted by the water flowing under the lateral moraine; the water has worn channels with vertical walls in the ablation moraine in the process of formation and the underlying ice. The upper part of the walls consists of drift and the lower part of ice, but at times the water flows beneath the surface. As a result of the streaming of the water, the ice melts faster and the drift deposited in the holes and channels is sorted. And as a consequence of the simultaneous melting and washing process, an irregular ablation moraine forms in the place, the material of which is in spots unsorted surface drift directly freed from ice and in other spots washed drift.

Ablation moraine forms likewise in places where medial moraines reach the glacier margin. In fact, the medial moraine in front of Heinabergsjökull described in the foregoing (p. 49), which was classified among radial moraines, is composed at certain points of mounds and depressions lacking to such a degree in orientation that these sections might be regarded as ablation moraines, even though the formation as a whole has a distinct radial lengthwise orientation. The medial moraine terrain extending from the great end moraine of Breidamerkurjökull to the glacier is so broad, on the other hand, that it lacks any distinct longitudinal direction. This area, consisting of medial moraine, taken as a whole too, is accordingly most nearly an ablation moraine, the surface of which is irregular on account of mounds and hollows of different shapes. If the glacier were some day so disappear between the starting point of the medial moraine and the present ice margin, it would likewise form a broad radial moraine.

Ablation moraine occurs on the outlet glacier of Iceland mainly in places where lateral and medial moraines come to the ice margin. The ice covered by such moraines melts so slowly that it becomes detached from the part of the glacier in motion and turns stagnant. Elsewhere there is usually so little surface drift that it does not produce ablation moraine topography proper. An exception, however, consists of such points where a border crest occurs on the snout (p. 35). The part of the glacier behind it melts faster than the ice underneath the border crest and in sinking lower forms a fosse, onto which surface drift slides down both slopes. Inasmuch as meltwater collects in the fosse and erodes it, the border crest is gradually detached from the glacier proper to form an end moraine under which is buried slowly melting ice, which deforms the moraine. In the event that the border crest and the fosse are situated so high or on such a sloping base that it does not collect water, either ablation or ground moraine topography is created by drift at the fosse. More commonly, however, the fosse becomes submerged upon sinking down and in its place a sediment plain forms.

Kettle topography of the ablation moraine type forms also in areas where glacifluvial material has accumulated on top of the ice. In front of the visible snout of outlet glaciers there are at many points on the surface sorted sediments. When the ice melts underneath them the smooth surface of the sediment becomes deformed so as to resemble ablation moraine. Considering that the material of the formation is not a direct product of the action of the glacier but it has been transported and deposited by flowing water, the treatment of such accumulations will be taken up in connection with glacifluvial deposits.

Kettle topography resembling ablation moraine in morphology is produced also when ice blocks broken off the glacier margin melt away. Although the glaciers of Iceland generally terminate on land, kettle topography at times occurs in the frontal terrain. This is true e. g. of Skeidarárjökull, where the kettles are formed as a consequence of glacier outbursts (comp. Leiviskä 1928 p. 188, Nielsen 1937). The blocks of ice torn loose by flood waters and transported to the sandurs are buried in part or entirely under drift carried by water and melt underneath slowly. In place of the ice blocks there then form hollows inside the accumulations; the roofs eventually cave in, leaving holes. Other kettle holes have been open from the very beginning and their sides have been steep; but both types become funnel-shaped fairly rapidly when gravel slides in. If the ice blocks had been densely distributed, the holes are situated so close together that the level stretches between them vanish and they are separated by sharp-crested gravel ridges.

The kettle topography produced by melting ice blocks is met with also at the bottom of ice-dammed lakes. The basin of Gjánupsvatn emptied in July 1949. At the same time the margin of the glacier had become broken

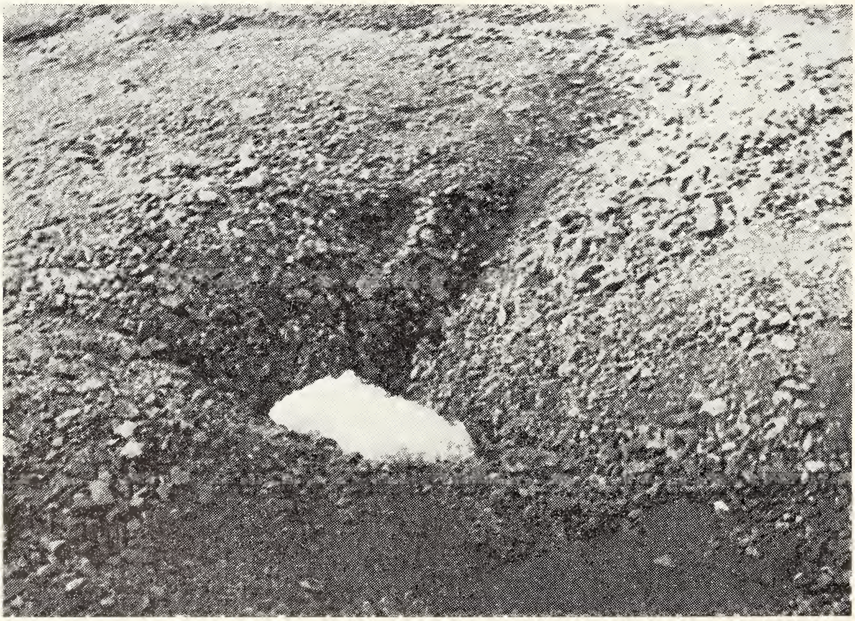

Fig. 23. A melting ice block in a funnel-shaped hole, which has been pressed into the bottom of the ice-dammed lake Gjánupsvatn when this ice block anchored. The ice remnant is about 50 cm in diameter.

up and the detached ice blocks had formed miniature ice-bergs. As the waterlevel fell, the icebergs caught on the bottom; and since the bottom was soft sediment or, in places, water-soaked till, it gave way beneath the weight of the blocks, forming depressions (Fig. 23). Melting blocks of ice were to be met with in such holes at the bottom of the vanished lake in great abundance. The ice blocks varied in size and the melting took place rapidly. Some of the holes were already empty. The holes created by the ice blocks measured at most 1 meter in depth and less than 5 meters in diameter. Inasmuch as the exposed lake bottom dried at the same time as the blocks melted, it appeared probable that the holes would remain at least until the waterline of the lake rose again to their level.

The kettle and knoll topography characteristic of ablation moraine thus forms in several different ways and in different accumulations. Ablation moraine proper in this topography is represented only by such formations as have originated upon stagnant parts of a glacier slowly melting from underneath overlying glacial drift. The amount of drift and its prevalence in the different part of the glacier determine in turn where the ablation moraine forms.

END MORAINES

By end moraine is meant a ridgelike accumulation formed in front of and parallel to a glacier margin as a result of either thrust or deposition action of the ice (comp. Flint 1947 p. 127). German researchers, such as Gripp (1938) and Woldstedt (1954) divide end moraines into two chief categories *(Stauchendmoränen* and *Satzendmoränen)* according to the manner of their creation. In North America end moraines are classified according to three types: push; dump and lodge moraines (Chamberlin 1894 b). Of these the push and dump moraines correspond to the main German divisions. Lodge moraines have originated out of basal load pushed by the ice margin and squeezed in front of it as a marginal ridge. In the German terminology such moraines fall under the heading of *Satzendmoränen.*

End moraines (also termed terminal moraines) are notable along the terminal margins of the southern outlets of Vatnajökull. As high and long ridges, generally visible from a considerable distance, they separate the glaciers from the coastal plain. Very well known are e. g. the great terminal moraines of Skeidarárjökull, Kviárjökull and Breidamerkurjökull, which were described already by Helland (1882), Keilhack (1883), Thoroddsen (1905—1906), Herrmann (1907) and Spethmann (1912) and which have subsequently been dealt with in numerous studies. Among the latter, noteworthy examples are the reports by Leiviskä (1928), Todtmann (1936, 1951 and 1952) and Hoppe (1953). According to these investigators, the end moraines in front of the southern outlets of Vatnajökull belong mainly to the class of *Stauchend*-moraines, though among them occurs also the structure of the *Satzend*-type. Leiviskä drew attention to the fact that under the basal till of the great end moraine along the eastern terminal margin of Breidamerkur there occurred folded peat layers (op. cit. p. 193). His explanation of the phenomenon was this: the accumulation was created by ice advancing over an earlier end moraine, folding its surface layer and then depositing a new moraine bed over the folded layers. The same kind of structure occurs also in the end moraine of Svinafellsjökull at Öræfi (Todtmann 1952, Hoppe 1953) as well as in the end moraine of Kviárjökull (Hoppe op. cit.). The occurrence of ground moraine on top of *Stauchend*-moraine need not, in Hoppe's opinion, signify a new thrust of the ice, however, but the ground moraine could have accumulated along thrust planes in the ice from uprisen basal load, which upon the melting of the ice has deposited as ground moraine on top of the push moraine (op. cit. p. 257).

Large end moraines occur also along the northern margin of Vatnajökull. The end moraine that formed in 1890 in front of Bruárjökull is, in the view of Spethmann (1909), Thorarinsson (1938 a) and Woldstedt (1939), mainly of the *Stauchend* type, but, as Woldstedt later stresses again (1954 pp. 101—102), it varies in nature greatly, even at short intervals. In one spot it consists of folded layers, in another it forms a broad zone of knolls and in a third just an irregular accumulation of boulders. It is thus hard to divide end moraines into different types on the basis of the evidence of special points, especially when there are few cross-sections and the moraines are situated so far from the ice margin that their mode of creation can no longer be observed. A couple of points, *i. e.* the termini of Hoffelsjökull and Svinafellsjökull at Öræfi, the glacier still extends into large end moraines, and at these points Hoppe has become convinced that the squeezing forth of basal till in the ice margin and transport in open crevasses play a major part in the formation of the end moraines (Hoppe 1953 p. 257). They would thereby be *Satzend*-moraines, or, according to American terminology, most nearly lodge moraines.

The margin of the eastern outlet of Hoffelsjökull extended in 1949 at several points to the end moraine, which had earlier (according to Thorarinsson about 1890) formed along the ice margin. The end moraine rises up from the surface of the sandur as a plantless rampart of stones, which hides from view the ice tongue, visible only from the summit of the moraine. The breadth of the end moraine on top is 50—70 m and its height, measured off the surface of the sandur, about 6—7 m, but on the proximal side the ground is lower and the height of the end moraine may be estimated at about 10 m, measured from the foot of the proximal slope. The rampart is almost symmetrical, but the slope varies so greatly that in places the distal side and in other places the proximal side is steeper. In those places where the slope of the proximal side is gentle, earth has obviously slid down from the end moraine ridge toward the glacier. This is quite comprehensible, as the proximal slope was generally wet, while the distal side and the top were dry. The summit of the end moraine was unexpectedly uneven and consisted of several small ridges. The ridges were linked together by lower sections in a chain, but often they ran parallel and merged. In between the small ridges lay elongated hollows, some of which contained water, though the bottoms were situated above the level of both the sandurs and the waters at the glacier margin. The surface of the small ridges, together with that of the distal side, was covered by uniform stony till, whereas there were fewer stones on the wet proximal side and the spaces in between consisted of normal till, which, however, was not pressed into as dense a mass as the ground moraine, for it could be shoveled like sand. There were two kinds of stones in the end moraine, some being angular and the others rounded (Fig. 19, Plate X). On the surface of both careful search would reveal fine scratches. In addition, the material of the proximal slope of the moraine yielded three mollusc shells, classified by Dr. Segerstråle of the Department of Zoology, Helsinki University, as belonging to the species *Saxicava artica* and *Leda permula*. Both are forms native to the Arctic Ocean. The same species have previously been found along the margin of Hoffelsjökull (Pjeturss 1907). Inasmuch as the *Leda permula* exists at present in water at least 30 m deep, its occurrences by Hoffelsjökull prove that the sea had extended up to the area nowadays covered by the glacier. Well-preserved fragments of birch roots and stems were also found in the proximal slope of the end moraine. Pieces of wood appeared in such abundance out of Hoffelsjökull that they were collected for firewood.

Underneath the surface the material of the end moraine proved to be loosely deposited till containing stones of various sizes. Three fabric analyses carried out on top of the end moraine at different points to a depth of 1 m yielded no certain orientation. It would appear that the till of the great end moraine along the terminal margin of Hoffelsjökull lacked orientation (comp. Hoppe 1953 p. 257). A sample taken at a depth of 1 m contained, in addition to mineral matter, some pollen and spores (*Betula* 4, *Pinus* 1, *Caryophyllaceae* 1, *Cyperaceae* 12, *Graminaceae* 4, *Myriophyllum alterniflorum* 1, unidentified 1 as well as *Polypodiaceae* spores 2, or a total of 24 pollens and 2 spores).

The material of the end moraine of Hoffelsjökull proved to be distinctly stratified in some places. The stratification was caused by the alternation of layers of coarse sand with others of fine sand. The borders between the layers were irregular and often the layers of coarse sand broke up and became divided into wedges, sunk into fine sand. The dip of the layers proceeded every which way. The structure of the moraine gave the impression that the sorted material had been sedimented elsewhere and been transported along with drift to the end moraine. In the main, however, the end moraine of Hoffelsjökull is unsorted, if loosely stratified till, which occurs on the surface of the moraine both on top and both sides. On the proximal side the surface consisting of it is wetter and less stony than elsewhere.

The snout of Hoffelsjökull formed in places an eaves-like projection over the base of the moraine. Blocks of ice falling off the snout had raised the mushy moraine and the overlying sediments into wall-like ridges, as related on p. 45. The confused and, in its detailed topography, varied structure thereby created was joined to the proximal part of the end moraine. Since the miniature ridges thus built up correspond in composition, structure and form to the ridges occurring on top of the end moraine, it is likely that both had evolved in the same way. The ridges and hollows in between them on the summit of the moraine would, accordingly, have originated through the falling down of ice fragments from the glacier margin, with the hollows indicating where the blocks had dropped and the ridges the accumulations of drift squeezed upward by them. Upon its happening, the ice margin would have had to extend as far as the end moraine. Nowadays the snout of the glacier is so low and is located so far from the end moraine that it works up the lower part of the proximal slope only at a few points.

West of the »calving» point the ice margin changes into a drift-covered border crest, behind which the surface of the glacier sinks down into a fosse (see p. 35). The layer of drift covering the border crest was about 50 cm thick and contained gravelly and sandy till. If there were more surface drift, it would protect the ice from melting even after the border crest had become detached from the glacier proper. At the site of the border crest there would then evolve an end moraine, which would contain ice melting in place and deforming the moraine. The thin drift covering over the present border crest would hardly suffice to bring about the creation of an end moraine. Possibly earlier, when the ice margin had been thicker and more active, it may have brought more drift on top of it, whereupon end moraines could have been built up also out of ablation moraines covering the margin.

At the margin of Hoffelsjökull there can be observed also a third way in which end moraine forms. East of the eaves-like projection mentioned in the foregoing the snout contains so little glacial drift that on the surface of the ice there are, in addition to the film of dirt, only little stones melted free of the ice here and there. The outlet terminates in a low, wedge-like snout, overhanging the basal till, which slopes downward toward the glacier. In front of the margin there occurred a symmetrical ridge about 100 m long and 1—0.5 m high. It was in contact with the ice edge at several points, but generally its crest was situated 1—3 m away from the margin. The ridge had evidently been built up out of basal till the winter before. It could well be a »push» moraine resulting from a slight advance of the glacier. In shape and size it resembled so-called annual moraines. The same kind of end moraine built of basal till occurred also along the terminal margin of Hoffelsjökull's western outlet, Svinafellsjökull. The crest of this moraine was situated at a distance of 8 m from the visible margin of the glacier. The guide, Leifur Gudmundsson, who knows Hoffelsjökull very well, reported that the margin had been situated at the point of the ridge the winter before. The grain size analysis (Fig. 19 p. 42) shows that the material consists of silt and clayey till, in addition to which the following pollens were found: 1 *Betula*, 1 *Cyperaceae* and 1 *Graminaceae*, as well as 1 *Polypodiaceae* spore.

The end moraines of Hoffelsjökull have, in the present author's view, been created out of material transported by the glacier. They therefore constitute »dump» moraine, with the reservation that the last-described minor ridge accumulations of basal till are »push» moraines.

Corresponding end moraines lie along the terminal margins of other ice tongues too. The large and irregular end moraine of Heinabergsjökull glacier is situated on the whole far from the ice margin and only at the northeastern end of the outlet does the ice extend up to the proximal slope of the moraine. The terminus of the glacier forms an arching projection at this point (Fig. 20, Plate X), overhanging the ground moraine a couple of meters below. As the material of the ground moraine consists of wet till, it is obvious that it will be deformed when the snout crashes down. The same process as at Hoffelsjökull is thus under way in respect to the proximal slope at hand. To be sure, the glacier has here diminished in size to such an extent that the margin has already been cut off from the end moraine and sunk down below the level of its summit. The end moraine must have developed here too as the result of stronger and more effective glacial action than at present.

Next to the preceding observation site the ice margin extends into the proximal side of the end moraine in such a way that the border crest rising above the fosse constitutes the proximal part of the moraine. Its ice is dark, hard and uncrevassed, and it forms at least three sharp-pointed, overlapping ice sheets slanting upward. The uppermost rises above the surface of the end moraine. The ice sheets thrusting up from underneath it terminate on a lower level and their ends are covered with drift, which also fills the cracks between the ice sheets inclined upstream and narrowing in the same direction. In the fosse between the marginal crest and the glacier the ice covers like a bridge water collected underneath the glacier. The ice has thus melted free of its base at the point of the fosse and the bridge formed by it is supported only at either end. If the bridge were to break from the effects of melting, the marginal crest would lose contact with the glacier proper and melt in place, creating kettle topography in the proximal part of the end moraine. If, again, the glacier were to become more active, it would push the snout of the outlet forward, whereupon this would plow the summit of the end moraine more vigorously than the ground on the proximal side of the moraine, because the ice is there detached from the base.

Apparently it is in the latter fashion, *i. e.* through oscillation of the thin margin of the glacier, that the end moraines along the bank of the Hanypá river, in front of the margin of Fláajökull, have originated. They comprise four successive ridges situated on the south side of the river a full half a kilometer from the present ice margin. The section eroded by the river shows how the end moraine ridges lie on top of the glacifluvial gravel beds. The nearly horizontal gravel strata continue without any deformation under the end moraine (Fig. 21, Plate XI). The border between the end moraine and the glacifluvial material is abrupt. The end moraine consists of relatively coarse, gravelly and sandy but plainly unsorted till (Fig. 19 p. 42). The orientation of the stones is weak, though small maxima are to be noted in the directions of 320° and 230° (Fig. 20 d, p. 44). The former corresponds to the direction of the apparent motion of the glacier, for the ice margin lies in this direction from the observation site, while the latter runs parallel to the longitudinal position of the ridge. A more distinct orientation maximum (Fig. 20 e, p. 44) occurs at the summit of the end moraine along the margin of Heinabergsjökull. The material is clayey till and the longitudinal axes of the stones in it lie parallel to the apparent direction of the glacier motion, for 40 % of the directions measured ranged between 280 and 300 degrees, the margin of the ice being about 50 m from the observation site in the 290° direction.

The end moraine of Breidamerkurjökull forms a ridge chain nearly 30 km long and 1—2 km broad. In places streams have eroded channels through the end moraine. Their gravel beds slope seaward between the hills. In the main, however, the end moraine forms a uniform crescent rising above its surroundings and separating the sandur from the irregular, lowlying and in many places water-filled ablation area. The surface of the sandur starting at the distal side of the end moraine is situated higher than the base level of the proximal side. The glacier snout is also below the level of the sandur. The glacier must have extended beyond its present margin and been thicker at the time the end moraine and proximal part of the sandur had formed. The visible margin of the glacier has by now become detached from the great end moraine at every point.

Along the eastern margin of the tongue the end moraine has broken up into hills and small ridges, which the dry beds of streams once flowing from the glacier separate from each other. According to Thoroddsen (1905—06) the erosion action took place in 1869, when the Vedurá river overflowed its banks and flooded the fields cultivated down the coastal plane. Subsequently the wind levelled out the end moraine hills, which from a distance resemble dunes (Fig. 24). The zone of knolls is about a kilometer broad and over five kilometers long. It forms a unified arc, in front of which begins an outwash plain terminating in the sea. The surface of the knolls consists of drift containing sorted, rounded stones. At least in some spots it is stratified and the layers dip upstream. They are loose and sandy, resembling the surface layer of the sandurs. A fabric analysis of such material was carried out on the summit of the easternmost end moraine hills yielded a broad but weak maximum between the directions 210°—255°. This corresponds by and large to the apparent course of the glacier during its advance, inasmuch as it is very nearly perpendicular to the orientation of the ridge (320°—325°).

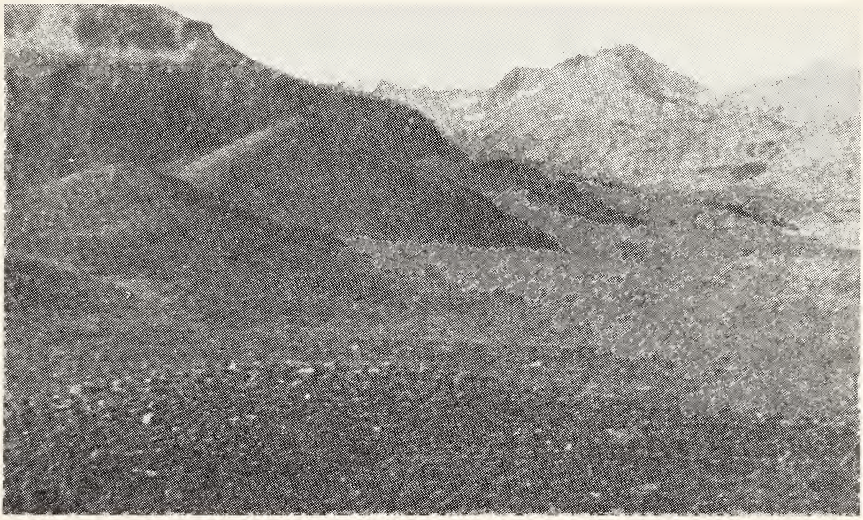

Fig. 24. The end moraine of Breidamerkurjökull appears in its eastern part as a group of separate, dune-like hills. The glacier left.

By and large the frontal moraine of Breidamerkurjökull is a coherent, broad and lofty group of ridges, whose parallel crests alternate with the long and narrow depressions in between. The sides of the frontal moraine are fairly even and slope about equally steeply both toward the sandur and the glacier. In places the distal side, the lower slope of which forms a distinct angle with the sandur, is steeper than the proximal side, which runs directly into the ablation area. Moreover, the distal slope is drier and has more vegetation than the proximal slope, which is sheltered from the winds. The middle part of the frontal moraine consists of several ridges, the summits of which ascend to different levels. The structure of such a ridge becomes evident in the old river bed, which intersects the end moraine about six kilometers from its eastern end. Although the section has collapsed, primary strata could be observed in them. In both sections they began at the upper slope of the distal part of the moraine ridge and inclined upstream (Fig. 22, Plate XI), but their slope varied and many of the layers were weakly folded. In the northern wall could be seen the thin, folded peat layers photographed by Leiviskä (1928) and Todtmann (1936) and in the southern wall too could be distinguished corresponding pieces of grass-peat in the collapsed gravel. The intersection consisted almost altogether of sorted sediments, for only at the summit of the ridge and in the section continuing from it to the upper part of the proximal slope there occurred a layer of till 2—3 meters thick. Many of its stones were cobbles, but their surfaces were scratched and the mass between the stones was dense clayey till, revealing a press structure (Fig. 23, Plate XII). A fabric analysis of the till yielded a clear maximum in the direction 345°—360°, paralleling the motion of the glacier. Inasmuch as the direction of the frontal moraine itself was 40°—220°, the orientation of the till produced a 40°—55° angle with the moraine. Underneath the till of the frontal moraine there occurred densely packed both fine and coarse sandy sediments. Such layers were met with under, between and on top of the peat beds (Fig. 24, Plate XII). Toward the surface their material became coarser, and in the upper part of the southern wall it was covered with a loose gravel of roundstones, extending up to the surface of the ground and forming there an esker-like ridge in the direction 40°—220°. A fabric analysis (Fig. 20 f, p 44) carried out on this ridge gave a distinct maximum parallel to the direction of the glacier motion. (The glacier was situated in the direction 325° from the observation site.) Since the material was so plainly oriented, it must be considered ground moraine, though it is not as dense as on the northern side of the river bed. Apparently the glacier had advanced on top of the »push» moraine; and it had accumulated earlier, transporting basal load there, as Leiviskä (1928) and Todtmann (1936) have explained.

The frontal moraine is intersected by the Jökulsá river, starting from the glacier. The fresh section of the eastern bank of the river showed that the moraine at this point consisted of sorted material, in which the stones were rounded pebbles less than 10 cm in diameter. Between the layers of stone there occurs sand beds. Both of them slope downstream from the glacier. The layers receive this slope already on the proximal side of the ridge and continue in the same direction throughout the entire 20—30 meter broad ridge. The slopes of the ridge are steeper and intersect the stratification, possibly being erosion forms.

On the western side of Jökulsá the frontal moraine spreads out and rises higher. It reaches its greatest magnitude at the point where the medial moraine meets the terminus of the outlet. The coarse material of this medial moraine, with its large boulders, covers the frontal area up to the distal side of the end moraine.

The Hrutá river starting from the western part of Breidamerkurjökull has eroded a channel for itself, like Jökulsá, through the frontal moraine. The section from the eastern bank of the river (Fig. 25) shows the frontal moraine to be composed of 3 parts: at the bottom an even-surfaced basal till making up the base of the whole ridge, a sediment overlying it in an undisturbed position and consisting of almost horizontal strata, and uppermost a till layer covering the sediments at two sites. One of these forms the proximal slope of the ridge and joins toward its bottom the till below and the other appears as a separate occurrence at the upper part of the distal slope. On account of the strong current the author could examine the occurrence only from the opposite bank of the river. From there one gained the impression that the ice had advanced over the place twice and that between the two advances there had been a peaceful sedimentation phase. In the sediment could be seen about 0.6 meters from the bottom a light yellowish ash layer in an undisturbed position. The later advance of the glacier had worn part of the sediment away but had not deformed the remaining part. Although the ridge in cross-section and orientation is a frontal moraine, it should perhaps more correctly be regarded as an

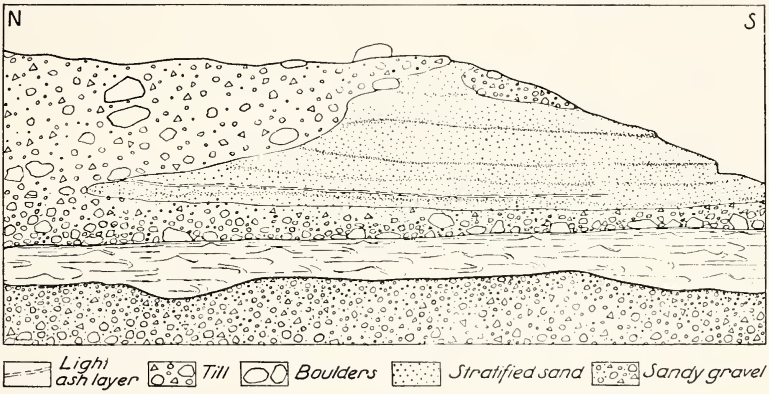

Fig. 25. A section of an end moraine at the river Hrutá in front of the western part of Skeidarárjökull. The glacier has moved from the north (left) carrying till on stratified sand without disturbing its strata. The section is c:a 5 m high. Drawn after a picture taken by the author.

erosion form, because only its proximal slope had gained its shape (and substance) as a result of direct glacier action.

On the hills on the western side of the Hrutá river there occur several successive moraine ridges, which run parallel in winding bands 100—300 meters long. They are 2—4 meters high and 10—15 meters broad at the base. The ridges are oriented southeast by northwest and they are situated on a slope slightly rising in the WSW-direction. At numerous points the northeastern slopes of the ridges were covered with blocks, whereas the southwestern slopes were sandy. These ridges were of the so-called annual moraine type, though exceeding in size ordinary annual moraines. Since the glacier had retreated long before, the manner of their origin could not be judged solely on the basis of the moraines. The topographic conditions, however, make it certain that the stony northeastern sides are proximal and the sandy sides distal slopes. On the other hand, it remains undecided, whether the ridges represent frontal or lateral moraines. The occurrence of glacial striae in the compass direction 325°—330°, about one kilometer southward from the end moraines, at Kvísker, strengthens the case for lateral moraines. This interpretation is also supported by the fact that on the eastern side of Hoffelsjökull by Gjánupsvatn, there occurs on the mountain slope the same type of lateral moraine, which indicates the position of the glacier flank in 1890 (Thorarinsson 1937 a p. 192).

Kviárjökull, situated on the western side of Breidamerkurjökull, descends in a form of an elongated and narrow valley glacier the south-

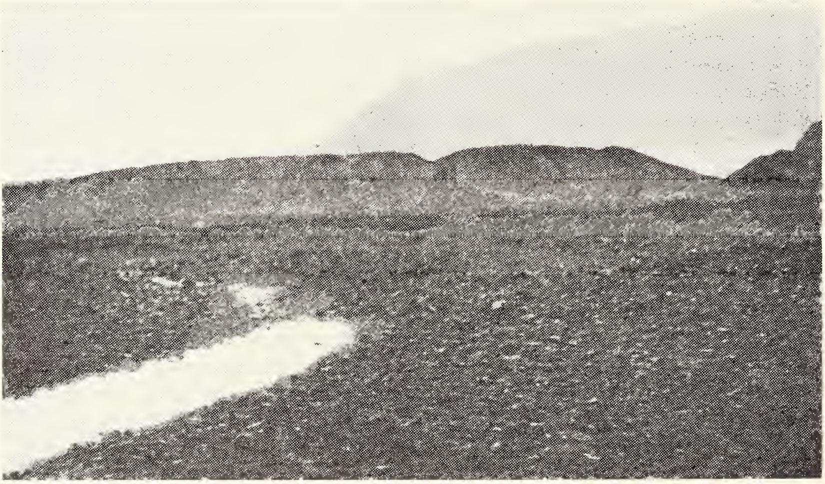

Fig. 26. The high terminal moraine of Kviárjökull seen from the east. In foreground, a ground water pool (light spot) on the sandur forms the start of a clear-water brook.

eastern slope of Öræfajökull to the coastal plain. This glacier has become known for its great end moraines (Thoroddsen 1905—06, Spethmann 1912, Todtmann 1936 and 1951, Hoppe 1953), which rise up from the low-lying coastal plain to heights of more than a hundred meters (Fig. 26). The frontal moraine is cut by a river starting from the glacier. The distal slope of the frontal moraine as well as the upper part of the sandur bordering it is sandy and covered with moss. Here and there on the distal slope are isolated blocks, some of which are very large (e. g. one measuring an estimated 7 × 7 × 7 meters) and they appear to have rolled down from the summit of the frontal moraine. The slope facing the glacier was bare and somewhat steeper than the distal slope, and traces of earth flow were to be observed in it. Also water flowing from the moraine had worn furrows into the proximal slope. This slope sinks lower without any clear boundary to become an ablation moraine, whose depressions are filled with water. Ice could be seen here and there and the ablation moraine gradually altered into broken-up ice covered with drift. The ice had turned stagnant by 1931 and the margin of the living glacier had forced up a transversal moraine on top of the dead ice (Todtmann 1936).

High frontal moraines like the end moraine of Kviárjökull occur in front of other hanging glaciers flowing down from Öræfajökull, such as Rotárfalljökull and Falljökull. The frontal moraines of the broad and gently sloping outlet glaciers are lower than the foregoing and form irregular groups of knolls. Their belt separates the sandur from the ablation moraine. Broad ridge and knoll zones occur along the margins of e. g. Svinafellsjökull and Skaftafellsjökull.

In front of Morsárjökull there are two frontal moraines crossing the valley. The outer one is of greater dimensions and more irregular than the inner one. The outer ridge has several short crescents, the convex sides of which face away from the glacier. The slopes of the furrow-like depressions between the crescents are asymmetrical in such a way that the side facing the glacier is generally steeper than the one facing away. The material of the crescents consists of fairly dense till. The longitudinal axes of the stones lie parallel to the motion of the glacier. This shows that the crescents were formed out of load directed by the motion of the glacier. They could have been thought to originate by the ice's having brought several sheets of basal load to the end moraine. These sheets overlap obliquely in layers sloping backward. The bed structure is visible also in the river bank at the northern part of the end moraine. It had, however, collapsed to such an extent that it was impossible to determine whether the whole end moraine had been composed of till or whether the till constituted only a surface layer. Layers of glacial drift rising obliquely upward are also to be seen at the low glacier margin sinking into meltwater and already drawn back from the frontal moraine (Fig. 25, Plate XIII). When the ice melts from in between the drift layers a sheet structure results also in the proximal side of the frontal moraine.

The frontal moraine of Skeidarárjökull is situated like the aforementioned between the sandur and the ablation moraine. The frontal moraine here consists of several sections separated by flat-bottomed gravel beds joined to the sandur. The short ridges of the frontal moraine running parallel to the ice margin are primarily irregular of surface (comp. Leiviskä 1928 p. 192), but in the view of the present author their shapes are also results of secondary deformation. Thus, the wind has both eroded and in places levelled out the original surface form. Glacifluvial streams have had an even greater effect on the frontal moraine. The often recurring glacier outbursts have broken the frontal moraine at many points. Likewise the present glacifluvial stream erodes it. E. g. the Kvistár, the middle one of the three streams starting from Skeidarárjökull, has changed its bed and worn away the frontal moraine on each side of the stream. In the same way the western part of the crescent has eroded into mounds called »Sandgigurs». The easternmost branch of the Nupsvötn has cut through one of these mounds. The section discloses the proximal part and middle of the ridge to be till rich in boulders. Near its upper surface there occurs in the till a fragment of a sandy layer about 20 cm thick, covered with boulders. In the distal part of the ridge the till is buried in sand, which forms the distal side of the frontal moraine. It joins the surface of the sandur as an embankment. The sandur layers continue under the sand up to the till. The border between the till and sand slopes sharply and more steeply than the distal side up to the level of the waterline of the stream. The formation has thus been built up of two different kinds of material, till and sand. They appear to have formed in such a way that the ridge of till accumulated first and only after that was the sand deposited on its distal side. Judging from the direction of slope of the layers, the sand had been deposited by water streaming from the glacier over the frontal moraine. Apparently, subsequent to the accumulation of the moraine, there had taken place such vigorous glacifluvial action that the end moraine was submerged. The water flushed the surface layer of the moraine and caused a sandy margin to accumulate along its distal side. The moraine evidently constituted at this point such a firm obstruction to the erosive action of water that the flood was unable to destroy it, though elsewhere it succeeded in breaking up the end moraine crescent, insofar as such a ridge ever occurred unbrokenly. As the end moraines of Skeidarárjökull have been subjected to greater erosion than the other tongues of Vatnajökull, it is natural that in its marginal moraine there should occur deviating forms and structure.

According to the view of the present author, the great end moraines in the frontal terrain of Vatnajökull's outlets originate during a more active stage in the life of the glacier than the present. The ice tongues extended farther at that time and were thicker than now. The snouts of the outlets have then folded the accumulations in the frontal terrain by pushing, and finally the glacier had advanced up to the »push» moraines thereby created, bringing basal till onto their surface. Furthermore, the observations made at Morsárjökull indicate that at least in places basal till had risen within the ice obliquely already at the time when the glacier extended as far as the end moraine.

In some spots the glacier terminus still extends up to the proximal part of the end moraine. The end of the snout usually melted free from its base forms an overhanging projection roughly in the shape of the pointed end of a ski, from which fragments of ice fall into the mushy till (p. 45). From the imprints made by the falling chunks and the till squeezed to the side there form hollows and miniature ridges in the proximal part of the end moraine. Besides these, only recessional moraines accumulating apparently in winter and the *Absatz*-moraines constructed through the detachment of border crests are in process of formation during the present stage. The creation of end moraines under current circumstances is rare, however, for the ice margin thins down so rapidly that it loses its power of motion and turns stagnant. Between the stagnant part and the living ice beyond, end moraines still form even at present by the edge of the live ice, thrusting over the dead ice and lifting basal till over it to produce a transversal moraine. When the ice melts from under the transversal moraine, it becomes deformed into a knolly and hollow-studded end moraine.

In general, the topography of the end moraines of Iceland is irregular, for even the large arcuate end moraines have formed out of short ridge mounds and the hollows produced by dead ice in between them. Moreover, streams have cut channels through the end moraines. From the stream sections it can be observed that the material of the end moraines is partly till and partly sediments from the frontal terrain. In addition, the end moraines contain superglacial drift mixed in with the till. Together they form inclined beds in the end moraine. A gently sloping distal side formed of stratified layers is a rarity in the end moraines of Iceland; and such a structure appears to have resulted only from exceptionally powerful glacifluvial action. Usually, the distal slope of an end moraine is clearly distinguishable from the sandur surface bordering on it, for the inclination of the latter is slighter.

ON THE GLACIER FLOW

The occurrence of glacial drift in the ice depends not only on the origin of the drift but also the movement of the glacier. On the basis of the drift, conclusions may thus be drawn regarding the mechanism whereby the drift is transported from its place of origin to the site of its occurrence. The migrating drift on top and inside the ice reflects the current behavior of the glacier, while on the basis of drift accumulations previously freed from the ice it is possible to reconstruct the behavior of the glacier at the time they were being formed.

Among moraines met with on the surface of glaciers, most interest is excited by medial moraines, which stand out distinctly from their surroundings (see p. 27). The longest of the medial moraines in Iceland, the great one on Breidamerkurjökull, indicates that the ice collects at the intake of the ice tongue in the center of the valley. The ice mass accumulating here continues its flow toward the terminus pretty uniformly. Only at the margin does the motion of the glacier begin to diverge, as evidenced by the spreading out of the medial moraine. The medial moraines starting from the nunataks situated farther down do not move toward the center during transit but run parallel to the winding course of the ice stream and remain the whole time at the same distance from the center line of the glacier. Inasmuch as medial moraines preserve the same shape for a long time, the movement of the ice tongue transporting the moraine must take place along the same courses, apparently determined by the topography. On the other hand, the fact arrests attention that medial moraines do not develop from nearly every nunatak. According to the present author's conception, the ice bordering on such nunataks is so thin that it no longer erodes their sides but melts free at bottom and becomes thinner from the effect of the heat stored up in the nunataks more rapidly than the ice farther away. As it wastes away both at the base and on the surface, the edge of the ice around a nunatak loses its erosive power.

The flanks of ice tongues likewise border in many places on warm mountain slopes. The edge of the glacier thins down in such cases, too, more rapidly than in the middle. In suitable spots between mountain and glacier there form ice-dammed lakes. The ice does not, however, fill their basins; but its main flow parallels the length of the outlet. Glaciers thus behave differently from flowing water, where pressure is distributed evenly in all directions (Flint 1947 p. 16). The transformation of lateral moraines into ablation moraines already in places where the middle part of the tongue is in motion as well as the process of detachment from the »living» ice offer evidence of the friction's growing so great along the flanks of the outlet that the main body of the glacier no longer is able to drag along the passive marginal ice. Between them there open up, as a result of the tension, crevasses, which separate the stagnant ice from the moving glacier.

The volcanic ash layers and dirt bands occur on the glacier in corresponding fashion: they form thin parallel bands that ascend forward between sheets of ice. The layers have not spread out nor scattered during glacier transport. This indicates that at least the rigid crust of the glacier moves as a sheet and its component parts advance at the same velocity. It also explains the fact that the material of the dirt bands does not undergo wear during transport but that e. g. the pollens contained in it remain intact and recognizable. Likewise the stones caught in the ice preserve their original form. The dirt bands and accumulations of ash travel to the ice margin in such a way that their flanks lag behind the center and turn parallel to the sides of the tongue.

The transversal moraines occurring in the terminal part of an outlet resemble in appearance dirt bands and ash layers, proving that the movement of the glacier is here inclined upward. The material of transversal moraines is, in many cases at least, thawed-out, mushy till, which is squeezed out onto the surface through transversal crevasses. Such a crevasse evidently acts as a shear plane (Philipp 1920, Seligman 1943). Till also may rise under hydrodynamic pressure (Nicols and Miller 1951 p. 278). Transversal moraines appear most distinctly along the border between moving and stagnant ice, but more commonly the snout of an outlet is divided into numerous successive sheets of ice separated from each other by drift bands. The movement of the glacier ends then by advancing along an upward inclination at a rate diminishing sheet by sheet toward the terminus. The friction in the basal part of the glacier must accordingly be greater than in the corresponding part of the surface layer, inasmuch as the ice at the bottom does not advance in the snout at the same rate as the ice layers overlying it. Since also the stones occurring in the sheets of ice and the hollows left in the ice walls upon their falling out are so situated that their longitudinal axes lie forward at an upward gradient, movement in this direction has taken place within the sheets too. The observations of the present author do not suffice to make it clear whether this motion is intergranular or whether it takes place also within the ice crystals (comp. Rigsby 1951, Sharp 1954), but they suffice to prove, in any event, that movement has not confined to the shear planes but occured within the ice as well. Motion directed along an upward gradient occurs in the snouts of outlets terminating as valley glaciers on a rock floor sloping downstream. The same kind of glacier movement may be noted also on the snout of Breidamerkurjökull, which, according to seismic soundings is situated on nearly an even floor. As Breidamerkurjökull forms an expanded foot at the coastal plain, an upwards-inclined movement of the ice occurs likewise at the margin of a Piedmont-type glacier. Upwards-inclined sheet movement has been described from the margin of Skeidarárjökull too (Todtmann 1936), where the snout has been observed to lie higher than the glacier bed behind it (Thorarinsson 1939 b p. 205). Thrust planes are met with, furthermore, in the terminal zones of Hoffelsjökull and Heinabergsjökull, where the border of the tongue rises uphill to the proximal slope of the end moraine. The obstructed flow thus occurs in the termini of the southern outlets of Vatnajökull regardless of the slope of the floor.

This type of movement is not characteristic of the southern margin of Vatnajökull alone, but it is encountered elsewhere in Iceland, too. As shown in Fig. 9 p. 23, the drift bands incline upwards on the ice cliff of Langjökull terminating in deep water of Hagavatn (Wright 1935). Transversal moraine bands occur also on the snout of the outlet terminating in the valley on the eastern side of Hagavatn, which snout lies on either a flat bed or one sloping gently downstream. Transversal moraines are to be observed even more distinctly on the surface of the broad lobes terminating in the highland of the northern margin of Vatnajökull (Spethmann 1909, Woldstedt 1937, 1939, Hoppe 1953). According to Hoppe's observations (op. cit.), accumulations like this occur especially on Bruárjökull, which resembles in shape the margin of an ice sheet terminating on flat ground.

Marginal ice thus moves along upthrust planes of the outlets, expanded feet and broad ice lobes of Iceland. Inasmuch as the slope of the floor does not seem to affect the manner of motion of the ice margin, it is evident that this motion results from the general behavior of the present glaciers of Iceland. In consequence of it, basal load rises within the ice in the shape of bands until it forms on the surface moraine bands running parallel with the margin. On Hoffelsjökull, Svinafellsjökull, Breidamerkurjökull and the ice tongue of Langjökull on the eastern side of Hagavatn the transversal moraines terminate at an elevation of less than 10 m from the terminus. On the ice cliff terminating in the waters of Hagavatn, they rise higher from the ice floor — an estimated 30 m. Likewise, at the margin of great Skeidarárjökull the moraine beds extended up to the surface in the 40 m-high ice cliff. On Bruárjökull the ground moraine at times travels to a height of several dozen meters (Hoppe op. cit. p. 248). As the transversal moraines usually come to an end at a distance of 100—200 m from the ice margin and as the strata carrying basal load dip backward, motion of sheets of ice reaching all the way to the bottom takes place only in the outermost part of the glacier. This is observable from the fact too that the crevasses of Hoffelsjökull extended below the level of the snout and that no drift was to be seen in their walls. The rising up of moraine onto the surface of the glacier causes in turn a retardation in the melting of the ice covered by it and the formation of border crests paralleling the margin. Under certain conditions, such as when a tongue has extended far below the climatic snowline and melts rapidly on its surface, a fosse develops behind the ice underlying the border crest; this eventually causes the border zone to become detached from the glacier proper. Thus the mechanism governing the motion of the glacier margin promotes the formation of stagnant marginal ice as well as creating irregular ablation and end moraine topography.

The accumulations of glacial drift that occur outside the ice margin, having lost contact with the glacier, afford an opportunity to investigate whether the aforedescribed movement mechanism of the glacier margin is characteristic only of the present stage of melting or whether the same kind of activity can be demonstrated to have taken place also earlier, when the glaciers of Iceland were larger than at present.

The thinning of the southern outlets of Vatnajökull has taken place so rapidly in recent decades that even the observations of Leiviskä (1928), Spethmann (1912) and Thoroddsen (1905—06) bring out differences in the ice margin as compared with the present. The ice tongues extended in their day to the great end moraines and the snouts were steeper and thicker than nowadays. The surface of the ice margin was in numerous places covered with surface moraine to such an extent that it was sometimes difficult to determine where the true margin of the glacier was situated. The moraine had, as Thoroddsen (op. cit.) mentions, risen to the surface of the glacier in layers inclined up toward the margin. Leiviskä (op. cit.), again, emphasizes the importance of englacial load as the carrier of the material of both eskers and end moraines, pointing to the observations made by him in Iceland. From the foregoing and other accounts dating back to the same period, it may be judged that obstructed flow had taken place also during the time when the tongues were larger than they are now. Simultaneously, there formed in front of the tongues large end moraines, the topography of which reveals that dead ice was buried underneath. Their formation in front of glaciers terminating on land can hardly be understood otherwise than by conceiving the ice margin as having operated in the way herein depicted. The structure of the end moraine of Morsárjökull, in particular, supports the conception of the ice moving obliquely upward. If this motion for some reason becomes stronger, the terminus of the ice tongue glides easily, like a ski, over the accumulations in the frontal terrain. Those accumulations standing in the way of the advancing snout are thereupon chiefly subjected to erosion. Push moraines are the result. It is from such material that the cores of the great end moraines are produced in front of both the southern tongues and the broad ice lobes terminating in the interior to the north (Spethmann 1912, Thorarinsson 1938 p. 495, Woldstedt 1954 p. 31). A marginal moraine of this kind came into existence along the northern edge of Vatnajökull e. g. in 1890, when the margin of Bruárjökull suddenly advanced. The advancing margin of the glacier thus creates an obstruction in its path and surmounts it.

Such behavior of the glacier margin is evidenced at numerous points. The end moraines appear on top of undisturbed layers at the Hanypá (p. 62) and Hrutá rivers (p. 65). The erosion of the floor by the ice margin has been extremely slight also in front of Heinabergsjökull, where the ice has advanced over the ground moraine and eroded its uppermost stones in the direction of the movement (p. 45). In the frontal terrain of Langjökull on the eastern side of Hagavatn the glacier has advanced across the same spot twice, without the later advance having erased the traces of the earlier erosion (p. 47). Even in the moraine the orientation of the stones caused by the earlier advance remains.

The observations presented in the foregoing regarding the transportation of glacial drift on the surface on the ice and within it show that the drift travels at the same velocity as the ice mass carrying it. The drift exposed from the bottom of the glacier reflects other conditions. It forms load in the basal part of the ice, increasing in quantity toward the bottom. This load is oriented in the ice of Hoffelsjökull and Breidamerkurjökull in such a way that the longitudinal axes of the stones are parallel to the ice movement. To judge from this, internal movement takes place also about the base of a glacier, during which its parts shift in relation to each other. There is only a slight amount of basal load in the bottom layer of marginal ice, for the main part of it melts free of the glacier earlier, forming ground moraine. This is pressed tight underneath the ice and preserves the orientation given it during the glacier motion even after the ice margin has advanced over it once more. The transportation of basal load thus takes place underneath the thick ice mass behind the margin. Without knowledge of the petrology of the glacier floor, it is impossible to judge how far the basal load has traveled during its transport, how rapidly it has eroded and where it has started to become detached from the ice carrying it. It is apparent that the accumulation of basal load depends both on the friction prevailing on the bottom and the formation and location of meltwaters. Inasmuch as both factors vary underneath the ice at different points, basal load accumulates irregularly during transport. The main erosion and transport of a glacier thus occur subsurficially and cannot be observed except by applying very special methods. Conclusions concerning this movement mechanism have been reached on the basis of the erosion forms of the rock bed as well as glaciological observations, and they have led to numerous theories of ice flow (see e. g. Sharp 1954).

From the erosion forms of the rock bed, the basal part of a glacier has been thought to move like a plastic mass (e. g. Demorest 1937, 1938, 1939, 1942). The theory has been supported by the evidence of so many investigations that it may be regarded as having been demonstrated to correspond to the facts. Moreover, observations have been made in Iceland (Hoppe 1953), which show the movement of a glacier to follow the microrelief of the floor. The present author clearly noted the same phenomenon in the rocks of Svinahryggur recently exposed from Hoffelsjökull (Fig. 26, Plate XIII). They form several less than 1 m-high »roches moutonnées», on the surfaces of which glacial striae are clearly visible. Even 0.5 m-high knobs of smooth outcrops divide the striations into two systems, turning around each side of the rocks. On the lee side they come together to form parallel striae again. The motion of the base of the glacier thus seems to have molded the detailed topography of the base and become adapted to its forms. Such motion calls for plastic ice. Since the very finest striae in the rocks have been produced under plastic ice and since they have remained fresh and unmarred, the rock no longer eroded after the glacier had lost its plasticity. According to Flint's conception (1947 p. 17), the rigid crust of a glacier extends to depths of 100 to 200 feet, below which added pressure causes the ice to flow together and close up all openings. Under certain conditions open tunnels have been encountered in ice even deeper (comp. Battle 1951 p. 563), which presupposes an even thicker rigid upper zone of ice.

Another proof of the plasticity of the ice eroding a rock bed is afforded by the gouges in the outcrop in front of Morsárjökull. They have eroded nearly vertically in the rock wall situated parallel to the movement of the glacier at the foot of the southeastern slope of the valley. The smooth surfaces of the gouges together with the thin and sharp striations in them show that ice impregnated with drift had pressed against the rock and eroded the gouges into their present form. The basal zone of the glacier has, in other words, striven powerfully in the narrow valley to expand sideward. There are no other signs of erosion at this point; but somewhat nearer the glacier, within the large end moraine, younger striae are also to be noted in the same vertical wall. They form a fan-shaped system of striae opening in the direction of the ice flow at a slant upward. Since in the end moraine there occur overlapping till sheets inclined in the same direction (p. 67), and since in the ice within it there can be seen drift bands that have risen at a slant upward (p. 67), these slanted striations must be regarded as evidence of obstructed flow occurring in the glacier margin. They are, accordingly, younger than the gouges worn by the aforesaid plastic ice. Erosion marks of two different ages occur also along the eastern side of Hagavatn, in front of Langjökull (p. 47).

The erosion marks on the rock bed accordingly demonstrate that the bottom zone of a glacier moves plastically and carries enough basal load to erode its base. The ice under such conditions must be subjected to strong pressure. Forms revealing such strong pressure, plastic ice and heavy erosion have either survived as the sole marks of glacial erosion in the rock bed or are clearly visible underneath erosion marks of later origin. This proves that a thinner mass of ice has operated at the later date, causing weaker erosion than the earlier thick ice; in fact, often the erosive action of the thin ice has not reached down as far as the rock floor at all.

The erosion marks on rock surfaces lead, together with a study of glacial drift, to the following conclusions:

1. Ice motion begins from the accumulation area either in such as way that the ice mass collects into the intake centers created by the topography and from there it presses out in the form of tongues or in such as way that the ice moves in a broad front outward from the firn area.
2. The movement of the upper zone is at first evidently directed at a slant downward, but in the ablation area it turns obliquely upward.
3. The surface ice of outlet glaciers advances toward the snout as a firm and rigid crust in such a way that its parts move along parallel winding flow lines. The movement is more rapid in the middle part of the tongue than at the edges or along the borders of nunataks.
4. If the glacier is sufficiently thick, its lowermost part flows like plastic material and erodes its base most. The motion in such cases takes place within the ice mass.
5. At points of obstruction the motion is retarded, while downhill in the direction of the flow it speeds up, whereupon the surface of the glacier is fractured.
6. The movement of the glacier in the marginal zone at an upward grade extends down to the base, where sheets of ice and the thrust planes between them lift basal load to the surface.
7. At the glacier margin the motion ceases in such a way that the sheets of ice separated by thrust planes slide over each other and this upward-inclined motion decreases from sheet to sheet toward the terminus.
8. At the frontal margin of the glacier and along the flanks of ice tongues there forms drift-covered stagnant ice, which participates in the movement of the glacier only in rare instances.

These conclusions are supported by measurements carried out at Hoffelsjökull (Thorarinsson 1939 b) and referred to on p. 14. Comparing his observations on ice flow with various theories, Thorarinsson concludes that in different parts of the same glacier appear different kinds of motion. At Skeidarárjökull he has seen the »Scherflächen» of Philipp (1920), at Kviárjökull »shear overlapping» (Todtmann 1936), and the meandering movement observed at Skeidarárjökull and Hoffelsjökull indicates plastic flow in certain parts at least, while in the more rapidly flowing parts of these glaciers there likewise occurs »block movement» or »Schollenartige Bewegung» (Finsterwalder 1929, 1931 and 1937).

The several theories presented in regard to glacier movement differ among themselves when dealing with the movement of ice on a horizontal and even floor. The basic conception behind the various theories may be illustrated, after Perutz (1953), by means of a rubber tube imagined to be forced vertically through the ice and subjected to deformation by the motion of the glacier. According to the main theories, the tube would behave in the ice in the manner shown in the illustration (Fig. 27).

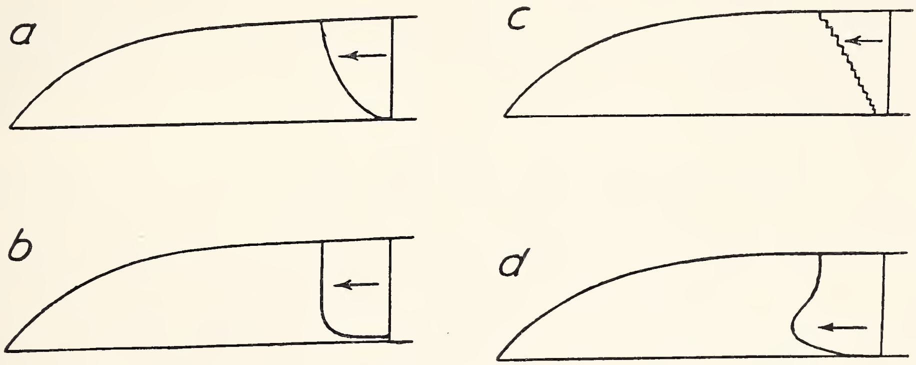

Fig. 27. Distribution of velocity along a vertical line through a glacier according to the main theories of ice flow (across on even floor): a. viscosity theory, b. plasticity theory, c. thrust plane theory, and d. extrusion flow theory. According to Perutz (1953).

According to the *viscosity theory* (Somigliana 1927), the tube representing the movement would be deformed into a parabola, wherein the part at the bed would remain stationary and the uppermost part advance most.

According to the *plasticity theory* (Hein 1885, Hess 1904), the lower part of the tube would be considerably bent while the upper part would remain vertical, inasmuch as ice must attain sufficient thickness before it can become plastic. The action of the glacier would thereupon be concentrated near the bed.

According to the *thrust plane theory* (Chamberlin 1894—95, Philipp 1920), the rubber tube behaves after the fashion of a broken line, the lower end of which moves least and the upper part most. The movement would take place along the shear planes within the ice in such a way that the uppermost layers slide forward more than the ones beneath them.

According to the *extrusion flow theory*, the tube is deformed in such a way that fairly far down there appears a bulge representing strong, fast flow within the glacier, while the movement in the surface zone and, especially, at bottom lags (Demorest 1937, 1939, 1942; Streiff-Becker 1938).

None of the theories herein presented can be considered as explaining the flow mechanism of a glacier, however, until it is backed up by precise measurements. This state of affairs has spurred researchers to elucidate the laws whereby glacial ice flows and to demonstrate in practice their validity. In recent years the theory of plasticity, in particular, has gained new supporters, for both the theoretical calculations (Glen 1952, Nye 1952, Körner 1954) and measurements carried out on glaciers (Perutz 1950, Haefli 1951, McCall 1952, Sharp 1953 a and b) consistently weight the scales down in favor of this theory. On the other hand, the plasticity theory has been criticized especially on account of the fact that the ice does not appear to turn plastic as a function of pressure, as presupposed in calculations, but the phenomenon is affected also by other factors, such as the formation of meltwater, the air content of the ice as well as the friction at the bottom (Sharp 1954).

The theoretical calculations together with observations made up to now thus lead to the conclusion that the surface crust of ice advancing over a flat floor is transported to a depth of about 100 m by the underlying plastic layer at approximately the same velocity but that lower down the motion slows down gradually toward the base, where the internal velocity of the glacier approaches zero. Such an explanation is in conflict, however, with the fact of erosion along the bed of the glacier, which remains unexplained. The floor of a glacier is seldom an evenly inclined plane. Usually the ice advances across a base made up of small basins alternating with steps. Accordingly, one might expect the topography of the floor to affect velocity in such a way that when the ice descends over a step it moves faster and becomes thinner whereas in the basins the ice thickens and simultaneously loses speed (Perutz 1953). According to Nye's calculations, the differences in the thickness of the ice cause the velocity distribution to be distorted from a straight line to an ellipse (op. cit.). This, in turn leads to stress and the production of shear planes intersecting the surface at a 45° angle. The sheets of ice sliding along these thrust planes erode the floor of the glacier, pick up the debris and carry the load up toward the surface. This would explain the erosive action along the glacier bed. It has been demonstrated through observation (Lewis 1949, Clark and Lewis 1951, McCall 1952) that the motion of a glacier takes place by rotation around an imaginary horizontal axis. Since these observations have been made in the ice cirques of Norway and the valley glaciers of Switzerland, generalizations cannot probably be made to the extent of also explaining the flow mechanism prevailing in the continental ice sheets. Only the future can show what importance the glacier rotation might have in explaining the floor erosion of ice sheets.

If the conclusions presented in the foregoing (p. 76) regarding the movement of glaciers in Iceland are compared with these theories, it must be affirmed that they do not suffice to decide according to what theory the glacial ice of Iceland moves. The erosion forms of the rock bed and the structure of the ground moraine show that the part of the glacier toward the base flows under pressure after the manner of a plastic mass, whereas the surface zone is rigid and brittle and rides on top of the flowing lower strata of ice. If the glacier thins, as happens both on the surface and at the base by melting, the rigid surface layer turns stagnant and moves only when pushed from behind.

In regard to the flow mechanism of the glacier margin, the different theories approach common ground. On the basis of the majority of them the motion of the glacier in the marginal zone is inclined upward, whether the margin in question be that of a valley glacier or continental ice sheet. This motion has been termed by the author obstructed flow, in conformity with most other researchers; and, as has been shown in the foregoing, the phenomenon is met with quite generally in the marginal zones of the glaciers of Iceland today. Examination of the accumulations of glacial drift and earlier observations made in Iceland prove that obstructed flow had taken place also in the past, when the glaciers of the island were more extensive than now and their margins extended to the great frontal moraines. As such a flow mechanism in an ice margin is independent of both the shape of the margin and the topography of its bed, it is evident that obstructed flow belongs to the mechanism of movement characteristic of the glacier types of Iceland. As a consequence of it, basal load rises in the marginal zone of a glacier to the surface of the ice, forming englacial load and increasing the amount of surface drift. Obstructed flow also causes the ice margin to erode its bed only slightly during an advance of the glacier. The erosion zone proper is situated farther back, underneath the thicker ice mass.

GLACIER DRAINAGE

The glacifluvial deposits produced as a result of action by water flowing from glaciers are an essential part of the glacial drift of Iceland. As the origin and appearance of these deposits depend directly on the action of water, it is necessary to consider the drainage of Icelandic glaciers before dealing with the glacifluvial deposits.

The abundance of water runoff the glaciers is determined by both the annual precipitation and the amount of unevaporated meltwater. Together they bring it about that from the southeastern and southern parts of Vatnajökull there flows into the sea such a volume of water that it has been compared to the amount emptying out of the Amazon river (Ahlmann 1936).

During glacier outbursts the runoff is multiplied many times over normal, causing short-term changes in glacier drainage. In the interior evaporation and less rainfall cause a decrease in the runoff to such an extent that the drainage of the glaciers in the interior differs from that along the southern margin of Vatnajökull. Common to both, however, is the lack of surface waters on the ice, for they disappear down into the crevasses and glacier mills, and do not appear before reaching the glacier margin. Only in exceptional cases does one meet with surface rivers, which flow over the ice in the ablation zone.

Ordinarily the water discharged from a glacier emerges only at the margin, where it collects in marginal lakes and pools of various sizes. From these the water then continues on its course, streaming powerful seaward. Thus the glacier drainage frequently takes place in three stages: the outflow from the ice mass, the formation of marginal lakes and the flow in braided rivers to the sea. Each stage affects glacifluvial drift in its own way. Accordingly, each of the stages is dealt with separately in the following.

OUTFLOW

The water discharged from a glacier may be classified in two categories, namely, surface water and subsurface water. The former includes the unevaporated water accumulating through rain and melting of the ice surface; it flows over the surface or along open crevasses to the ice margin. The subsurface water comprises that flowing both englacially and subglacially. Since it is often difficult to decide in what part of the glacier subsurficial water has flowed, no effort has been made in the following to break down the classification; the exact situation of the water within the glacier will be discussed in another connection.

A. Water discharged from the surface of a glacier. In spite of rapid melting and heavy precipitation, but little water collects into streams on the ice surface. Evidently the ice surface is too much broken up to favor the formation of surface rivers. In the upper part of the ablation area, where water begins to occur in abundance and where there are fewer crevasses, rivulets form on the surface of the ice; but they empty their waters eventually into crevasses. Only in rare cases does a surface river flow as far as the ice margin. Such surface streams and rivers flow through channels with a natural slope, which determine the rate of flow. The channels cut into the ice, and if drift collects on the bottom, the stream begins to meander. Such streams are not long-lived enough to develop any regular longitudinal profile, for they run dry when the temperature falls; and in winter the surface of the ice is waterless. The channel-systems of the previous summer disappear with the freeze-up and the following summer the water must wear down new channels. Thus in the summer of 1948 a surface river of unusual size flowed across the eastern part of Hagafellsjökull (see Hoppe 1950 p. 49), but by the following summer, when the present author visited the area, it had vanished. The river starting from the glacier in 1949 received most of its water from two snow-covered crevasses. Over the snowless and uncrevassed ice surface next to them there flowed only small streams with clear water. Similar streams occurred that summer almost everywhere on the southern ice tongues of Vatnajökull. Slight surface flow takes place on the glacier in the summertime also in the northern marginal zone of Vatnajökull (Spethmann 1909) and it appears to be general in the ablation areas of the Icelandic glaciers. The volume of water carried by surface streams is so small, however, that it has no appreciable importance in the drainage of the glaciers.

The crevasses on the ice surface naturally serve as collectors of surface water; but the water does not remain in them but vanishes from sight so fast that the crevasses are waterless to as great a depth as it is possible to see down into. The radial crevasses at the ice margin represent an exception in that there is usually water at their bottom. It has not, however, accumulated directly from surface flow but collects within the crevasses subsurficially. If the bottom of a crevasse is so high that the crevasse does not extend to the general water level, no water gathers in it.

B. Water discharged subsurficially. Most of the water of the Icelandic glaciers is discharged from subsurficial water systems, where it collects in different ways. Some apparently represents the water disappearing down into the crevasses and glacier pipes or tubes. Another part of it derives from the ice bordering nunataks and mountain slopes, where in addition to meltwater there collects rainwater flowing down the mountains; while a third part forms in the lower zone of the glacier. The masses of water released by subglacial eruptions likewise flow subsurficially to the ice margin. Thus Nielsen (1937) emphasizes that the water generated by the eruption of Grimsvötn flowed the entire 55-km distance from the eruption site to the margin of Skeidarárjökull subglacially.

Ordinarily the subsurface waters flow out of the mouths of tunnels along the margin from which water has previously been discharged with great force (Thoroddsen 1905—1906, Spethmann 1912, Todtmann 1936, Ahlmann 1939, Thorarinsson 1939). The mouths of the tunnels are situated for long stretches of time at the same points along the ice margin, but sometimes they become clogged, whereupon a new opening appears next to the old, blocked one. According to the observations of the present author, the mouths of the glacier tunnels situated along one or the other flank of an ice tongue near a point where the edge of the tongue began to curve free from the slope of the bordering mountain (Hoffelsjökull, Skeidarárjökull, etc.), or then at the center of the curved margin of the tongue (Breidamerkurjökull, Kviárjökull). The latter had formed at points where long medial moraines reach the glacier margin. E. g. the strongly flowing rivers, Jökulsá and Breidá, on the Breidamerkursandur originate at the site of the medial moraines. From all evidences, medial moraine promotes the generation of meltwater (comp. p. 00). The streams flowing out of the flanks of ice tongues derive, at least in certain places (Skaftafellsjökull's western margin and the eastern margin of Hagafellsjökull), from the water collecting under the lateral moraine at the point of contact between the glacier and the bordering mountain; in addition to meltwater it includes rain water streaming down the mountainsides. Underneath the glacier margin the water flows laterally along the edge of the valley, for in places the babble of water can be heard and in other places the brook comes into sight in the openings. The water soon vanishes into tunnels, out of the mouths of which it is discharged. A dry tunnel was to be seen in the summer of 1949 in the stagnant ice at the southeastern margin of Morsárjökull. Both ends of the tunnel were open and it had obviously formed part of an earlier waterway but had fallen into disuse when the water had found a new channel for itself. The tunnel had formed in clean, unbroken ice, and only its lowermost part extended down into the bottom layer of ice containing basal load. At the bottom of the tunnel there was no loose glacifluvial drift. The meltwater flowing along the flank of the glacier had thus, at least in the case of this particular tunnel, run englacially without penetrating down to the floor.

The stream flowing down into the lake of Hagavatn along the eastern flank of Langjökull ran out of a tunnel that had been worn through moraine-covered ice bordering on a mountainslope, which ice became clean in depth. The tunnel extended for a distance of only about 150 m, however, for the stream changed farther up into a smoothly flowing river. Its point of origin was the mouth of a new glacier tunnel, from which the water was discharged in a swift and even flow. The water did not fill either tunnel wholly, but considerable air space was left beneath their arched ceilings. It was not possible to ascertain on the spot what the base of the tunnel was like; but next to the lower tunnel could be seen another, empty one, situated a bit higher, which had previously been in use. The walls, ceiling and floor of this tunnel consisted of clean ice. The diameter of the tunnel was approximately 50 cm. It was open at its lower end and its floor rose gently upstream.

The lakes dammed up along the southern outlets of Vatnajökull add, upon discharging subsurficially, to the volume of water in the rivers originating in the glacier. The discharge flow does not follow the joint between glacier and bordering mountain, but e. g. the ice-dammed lakes along the eastern flank of Hoffelsjökull discharge their water via the Austurfljót river, which starts from the southwestern margin of the same glacier (Thorarinsson 1939 b). The water thus flows slantwise underneath the ice. In the past the Austurfljót ran out of the glacier tunnel in mighty cascades (Thorarinsson 1939 b p. 204). Pressure of equal power used to prevail also at the mouths of most of the glacier tunnels, as e. g. at the discharge point of the Jökulsá, flowing out of Breidamerkurjökull (Ahlmann 1936), and during glacier outbursts it has increased above normal. During glacier outbursts caused by volcanic eruptions, the ordinary water ways cannot accomodate the whole volume of water discharged from the ice, so new tunnel openings appear. For instance, during the Grimsvötn eruption, meltwater was discharged out of thirteen tunnels opening up in the glacier margin (Nielsen 1937). According to Ahlmann (1936), Nielsen (1937) and Thorarinsson (1939 b), the flow of water in the glaciers takes places subglacially. Inasmuch as the margins of at least Hoffelsjökull, Heinabergsjökull, Breidamerkurjökull and Skeidarárjökull are higher than the floor the water must flow uphill in order to emerge from under the ice. It must, in other words, flow under hydrostatic pressure. Observations of powerful spraying at the mouths of glacier tunnels seem to verify the thesis of a subglacial flow of water. Other evidence to support it can also be produced, such as e. g. the fact that meltwater erodes the glacier floor of Hoffelsjökull and Breidamerkurjökull (Thoroddsen 1905—06, Pjeturss 1907, Ahlmann 1937 b). Keld Milthers, who, as a member of Niels Nielsen's expedition, was the first to examine the mouths of glacier tunnels opened up by the glacier outburst at Skeidarárjökull and who entered a tunnel, related that the tunnel was, in truth, at the bed of the glacier and, consequently, had been opened up in the ice by subglacial flow of water (oral report).

All students of the glaciers of Iceland do not, however, agree with the thesis that meltwater flows within the ice along the bed under hydrostatic pressure. Leiviskä (1928) emphasizes that the pressure of the water discharged from glacier tunnels is insufficient to transport large amounts of gravel except during glacier outbursts. The same conclusion was reached by Todtmann in making observations in the summers of 1931 and 1934 of the conditions prevailing along the southern margin of Vatnajökull (1936). In studying the mouths of glacier tunnels she noted that the meltwater flowed from them in channels with a natural slope. Only in the event that the glacier breaks up or sinks down to block the stream does the water dam up and issue forth in powerful eddies. When the obstruction has been worn away, the slope of the stream is restored and the mouth of the tunnel begins to function regularly again. As the surface of the sandur and the ice margin bordering on it are situated above the level of the glacier bed, the meltwaters running out of tunnels down a natural gradient must, according to Todtmann's conception, flow englacially rather than subglacially. Todtmann further argues that the ice flow is retarded first in the basal part of the ice, over which the surface ice advances along shear planes. These planes guide the meltwater deriving from the glacier surface and aspiring downward to the margin instead of its penetrating through the rigid basal zone of the stagnated ice (op. cit. pp. 84—85).

The question of the manner of flow of subsurficial water is thus for the time being an open one. Speaking in favor of Todtmann's conception are the shear planes, the results of seismic sounding (see p. 11) — which show the expanded ice of Breidamerkurjökull to be so thick that the existence of subglacial tunnels at its base becomes doubtful — and the absence of pressure at the point of origin of the Jökulsá, though the glacier bed is still below the ice margin. It is also hard to imagine the meltwater of the medial moraine penetrating a mass of ice 500 m thick, flowing along the glacier floor and being discharged uphill without inducing hydrostatic pressure. The diminishing of pressure is no local or accidental phenomenon, for the mouths of the tunnels studied by the present author had changed in nature. The Austurfljót, which starts from Hoffelsjökull, formed a lake at the site of its emergence, which extended into crevasses in the ice margin and received water from the glacier so evenly that no precise point of discharge could be detected. The water flowed calmly forth also at the eastern margin of Breidamerkurjökull, from under Fláajökull and Heinabergsjökull. In front of the margin lakes dammed up by end moraines had formed at numerous points; the lakes continued under the ice, with the snout forming a roof over the water. Under the ice was 5—30 cm air space and the flow could be deteched only by watching the movement of objects floating on the surface of the water. The Swedish investigators Hjulström (1953, 1954 a and b) and Hoppe (1953) noted the same phenomenon in the summers of 1951 and 1952. The lakes forming in front of the southern tongues of Vatnajökull had remained and gathered the water running from the glacier. Hydrostatic pressure has thus, in the view of the present author, generally diminished in the subsurface streams of Iceland. As an illustrative example, be it mentioned that Jón Jonsson, who, as a member of Hjulström's expedition, studied Hoffelssandur, reported having found the young of sea salmon in Gjánupsvatn. The salmon had to swim 5—6 km upstream to reach the lake. Hjulström (1954 b p. 34) regards the formation of marginal lakes a consequence of the current climatic change.

The discharge of the waters of the Vesturfljót river, which originates in the ice of Svinafellsjökull, gives a good picture of how water at present issues forth from a glacier. Part of the water purled as little brooks over the glacier surface and part flowed into view as a somewhat larger but gentle stream from the mouth of a tunnel at the southeastern corner of the ice tongue. This was situated under the ice edge in such a way that the arched lower surface of the glacier formed the ceiling of the tunnel. The bottom of the tunnel consisted of soft and clayey moraine, which the flow eroded. About 15 m from the ice margin the volume of water in the stream increased many times over, for it was fed additional water from five Artesian springs welling up through its bottom, (Fig. 27, Plate XIV). These springs were situated in water-saturated till outside the visible border of the glacier. As the moraine was unable to obstruct the passage of water, there must have been a denser layer underneath through which the water under pressure forced itself into the river. A separating stratum consisted of ice buried in the moraine, the existence of which could be ascertained by prodding it with a rod. The true margin of the glacier thus extended beyond the visible border. This meant, at the same time, that the mouth of the tunnel at the border of the glacier constituted, in fact, the point of discharge of englacially flowing water.

In front of Heinabergsjökull the water came as a river flowing from Heinabergsdalur, on the northern side of the ice tongue. Judging from the dry beds, the river had earlier flowed past the terminus of the ice tongue seaward (comp. Thorarinsson 1939 c p. 219). In the summer of 1949 the river curved toward the terminus and flowed in the direction of the glacier margin, in places being bounded by the glacier margin and in other places vanishing under the ice. Finally, the flow grew weaker and the water formed marginal lakes between the ice and the end moraine. The volume of water in the stream increased gradually, for it appeared to collect the meltwater flowing from the glacier. The flow was so smooth that it could hardly be distinguished. No hydrostatic pressure occurred here either, therefore, but, on the contrary, the stream even flowed back under the ice.

The waters of the eastern margin of Breidamerkurjökull issued out of the mouth of a glacier tunnel. The surface of the glacier had sunk into a hollow at the spot. The water flowed gently out of an opening about 20 m wide, which had an even ceiling of ice 10—20 cm above the water level. The water was so turbid and the opening so hard to approach that it was not possible to test the depth of the stream or examine the bottom of it. The river received additional water from 5 or 6 little brooks flowing on the surface of the glacier and eroding furrows in the margin of the ice tongue. The volume of water brought by the brooks was insignificant compared with that discharged subglacially (perhaps 1/100). In front of the ice margin the river flowed through a steep-walled channel 2—3 m deep running across the gravel plain; it received additional water from Artesian springs welling up through the bottom of the river; this increased the volume of water by an estimated half. The stream became swifter over a distance of about 100 m and produced small rapids, the sills of which were formed by a hard and unbroken stretch of ice (Fig. 11, Plate VI). This stratum of ice, which the river was eroding, continued on each side of the stream bed underneath the gravel plain. Here, too, outside the visible terminus of the glacier, there was ice buried under glacial drift, ice that formed the true but invisible margin of the glacier somewhere on the proximal side of the end moraine. Thus, the aforedescribed opening must, in the present author's view, be situated here too between two sheets of ice and represent the point of discharge of the englacially flowing stream. The fountains occurring at the bottom of the stream, from which water spouted 20—30 cm above the surface, evidently received their pressure water from under a sheet of ice, as in front of Svinafellsjökull. The river running parallel to the ice margin soon disappeared from sight, for its flow became subsurficial — within its own outwash (and the ice underlying it). From here it emptied into the lake formed on the proximal side of the end moraine.

A lake collecting meltwater had formed in front of Breidamerkurjökull also at the source of the Jökulsá (Fig. 11 p. 29), where it had previously emerged from the mouth of the tunnel. Now the tunnel had vanished and the water seemed to flow into the lake along the ice margin from both directions. The margin of the glacier descended at a gentle incline into the lake and small rafts of ice broken off it floated on the water. A great abundance of meltwater collected at the ice margin now, too, but its flow did not become swift until it reached the neck of the Jökulsá river. The Jökulsá was crossed in 1949 with a small boat by rowing over the lake, whereas previously the river and the mouth of the glacier tunnel were circumvented across the ice (Leiviskä 1928).

The Skaftafellsá river starting from the W-margin of Skaftafellsjökull flowed along the edge of the ice tongue as a subsurficial lateral stream. It came into sight as a small brook from between the clean-surfaced tongue and the dead ice buried under the moraine alongside. The brook received additional water from another brook with clear water running off the glacier surface as well as from numerous short tributaries originating in the ablation zone. The latter flowed through 7—8 m-deep »canyons», whose vertical walls consisted of firm ice underlying surface moraine 1—2 m thick. The brook valleys joining together from different directions combined to form two river branches, separated by moraine-covered islands of ice; and these two branches represented the source of the Skaftafellsá river (Fig. 28). Pressure water was not met with in this locality either, for the river started from meltwaters flowing together from different directions. On

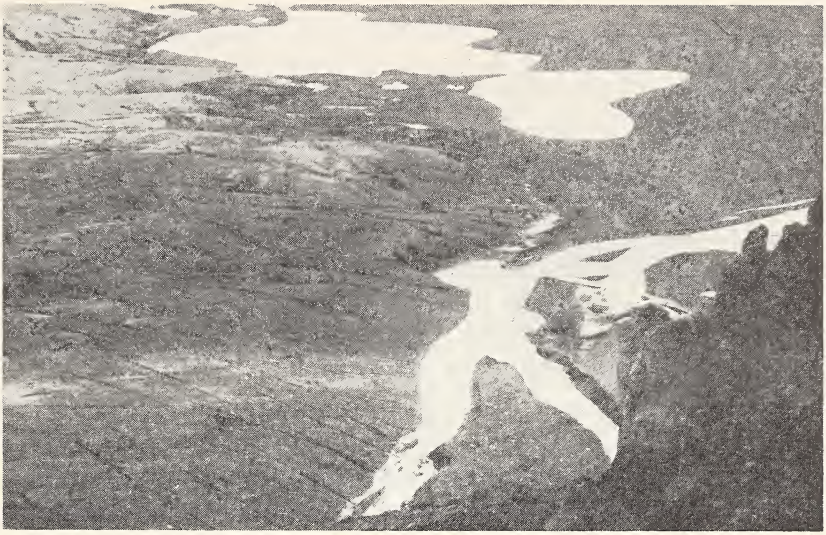

Fig. 28. The terminus of Skaftafellsjökull (left) seen from the bordering mountain. A lateral moraine and the source of the river Skaftafellsá separate the mountain from the glacier. Small push moraines appear in front of the glacier between the river and a marginal lake (in the background). Black stripes on the glacier surface are ice-cored drift ridges.

the spot one could not determine whether they flowed through ice-bottomed channels or whether they had cut channels through the ice.

Pressure water was met with in two places along the margin of Langjökull. At the eastern margin of Hagafellsjökull terminating in Hagavatn, near the mouth of the glacier tunnel mentioned previously, on p. 82, there occured on top of the clean-surfaced ice, which sank at a gentle incline into the lake, two fountains. The water spouting from them was clean and rose above the surface of the glacier. The other Artesian wells occurred at the margin of eastern Hagafellsjökull. At a distance of about 20 m from the visible margin of the glacier there had appeared on the ice two fountains, from which the water spouted to a height of about 30 cm. At both points so little Artesian water was discharged that it had to derive from slight, local pressure. There was no outflow point for subsurficial meltwater streams at either site. These two observation sites were the only ones where pressure water was obserwed to rise up from the clean surface of the glacier.

The observations described in the foregoing show that a large volume of water is discharged from the glaciers of Iceland, especially during the summers. It flows in the ice subsurficially, for only a slight amount of water reaches the ice margin on the surface of the glaciers. Under normal conditions subsurface water runs to a large extent through englacial tunnels.

Outside the visible ice margin pressure water is encountered in places where the true margin of the glaciers extends beyond the visible terminus. This water flows evidently subglacially.

The volume of subsurface water multiplies during glacier outbursts many times over, and the water is discharged under high pressure from the mouths of glacier tunnels. According to the conceptions of Nielsen (1937) and Thorarinsson (1939 b), the water at such times forms subglacial basal streams, which extend from the point of discharge to the ice margin.

MARGINAL LAKES

Water drained in various ways from a glacier usually collects outside the ice margin, forming lakes and ponds, from which new streams flow onward. Part of these accumulations of water are ice-dammed lakes and part have formed behind the end moraines or proximal sides of the sandurs.

Ice-dammed lakes occur along the flanks of southern outlet glaciers, but they are met with also along the margins of other Icelandic glaciers. They have usually formed in the »armpits» between ice margins and bordering mountains, in which both rain and meltwater accumulate. The amount of water in such a lake is likely at times to be quite considerable. Iceland's largest ice-dammed lake, Græналón, along the flank of Skeidarárjökull is 18 km² in area and contains an estimated 1 500 million m³ of water. And Hagavatn, bordering on the margin of Langjökull, was 11 km² and contained 150 to 200 million m³ of water before the formation of its present discharge (Thorarinsson 1939 b pp. 240 and 237). The volume of water in ice-dammed lakes varies tremendously, however, for many of the lakes have subsurficial outlets. According to calculations by Thorarinsson (op. cit. p. 227), the sill of lake Graenalón was overlain at the turn of the century by an ice mass 300 m thick. At the same time there was ice 225 m thick at the sill of lake Vatnsdalur, dammed by Heinabergsjökull, but by 1938 the glacier had wasted down to a thickness of 170 m at the same location (idem. p. 222).

When the ice-dammed lakes are at their maximum, they are so deep that ice barrier begins to float and calve, with the walls of the margin becoming quite steep. Upon a lake's suddenly emptying out, its bottom is exposed and the streams running down the mountains flow across it toward the glacier, until they vanish under the ice or begin to refill the basin. Such a stream flowed in the summer of 1949 across the dry bottom of the Efstafallsvatn at the eastern margin of Hoffelsjökull. The flow halted only upon reaching the glacier margin, where water began to accumulate.

The lake of Hagavatn, (Fig. 4 p. 12) discharges its water through a col situated outside the glacier, the sill of which regulates the water level. Such a col serves as the outlet until the water finds a new, lower sill under the ice, whereupon the outlet shifts to this point. At the same time the waterline sinks down to the level determined by the height of the new sill. The present outlet of Hagavatn developed in 1939, when the water in the lake melted or eroded the glacier so much that it was able to flow out over a sill 9.5 m lower than before underneath the ice. The outflow lasted 3 days, during which the ice margin wore away 400 m, exposing the point of discharge (Thorarinsson 1939 c). Since the outburst the surface of the lake has remained at the level fixed by the new sill. Only the erosion of the outflow sill and the changing of the seasons have caused alterations in it. In the summer the water level rises, for water accumulates at that season faster than the outlet is capable of discharging. In the winter the waterline descends, since less meltwater is produced and precipitation is mostly in the form of snow.

If the margin of the glacier terminating in the lake of Hagavatn is studied during the season of high water, it will be noted how the ice tongue descends gently into the lake (Fig. 18, Plate IX). Only in the middle part of the lake does the glacier form an ice cliff (Fig. 9 p. 23), extending 5—6 m above the waterline. The lake is evidently so deep at this point that the glacier begins to float there and its margin to calve. No calving bay has, however, resulted, for the ice margin remains straight. In 1934, *i. e.* before the last outburst of Hagavatn, an expedition from Cambridge University drew up a depth chart of the lake (Wright 1935), which shows that in front of the ice margin at that time there had been water to a depth of 140 feet (42 m). Since then the level of the lake has sunk nearly 10 m lower and the ice margin has retreated. As the lake, according to its sounding profile, deepens toward the glacier, the probability is that there is at present water to a depth of 30—40 m. This depth suffices therefore to lift the terminus of the ice tongue free from the base, provided that the steepness of the glacier margin really results from this cause. In studying Hvitarvatn, situated near Hagavatn and likewise bordering on Langjökull, and the glacier margin terminating in that lake, Nørvang (1937 p. 188) came to the conclusion that warm surface water hollows out the glacier and causes calving of the upper part of the margin. This conception is supported by observations made at the glacier margin during low water. The sides of the ice tongue, which slope gently down into the lake during high water extend down below the surface also during low water. The position of the high water is indicated on the ice by an erosion indentation below which the gradient of the ice surface is appreciably gentler than higher up, in the supra-aquatic zone(Fig. 29). In the ice cliff along the central part of Hagavatn there is to be seen during low water at numerous points along the summery waterline a deep trough, which surging of the surface has worn into the side of the glacier. The trough is roofed over by a long eaves-like projection of ice, which has broken off in places and left ice blocks floating in the water (Fig. 30). The glacier margin terminating in water thus wastes away during high water (in the summertime) along the waterline and the level immediately below it, whereupon the supra-aquatic part of the ice becomes steep.

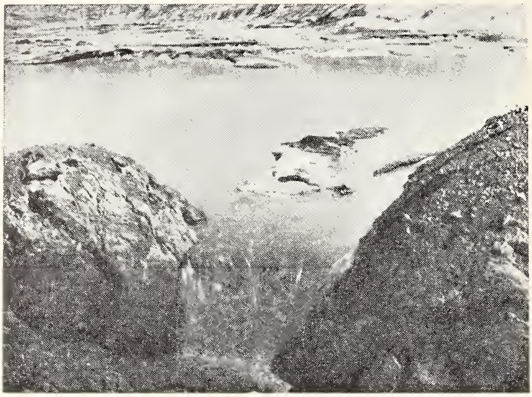

Fig. 29. A marginal lake, situated at the western side of Hoffelsjökull, is draining through a rock col (foreground). Before the outbrust the lake water melted a terrace into the ice (background) so that a thin sheet of ice has been left at the bottom of the lake. This ice becomes visible, forming islands behind the sill.

In deeper water the melting is slower and the ice there forms a thin, gently sloping sheet, which extends appreciably farther out than the margin at the waterline. Possibly this thin and therefore stagnant and probably sediment-covered sheet of ice is an obstruction to movement by the glacier and causes it to thrust upwards (see Fig. 9 p. 23).

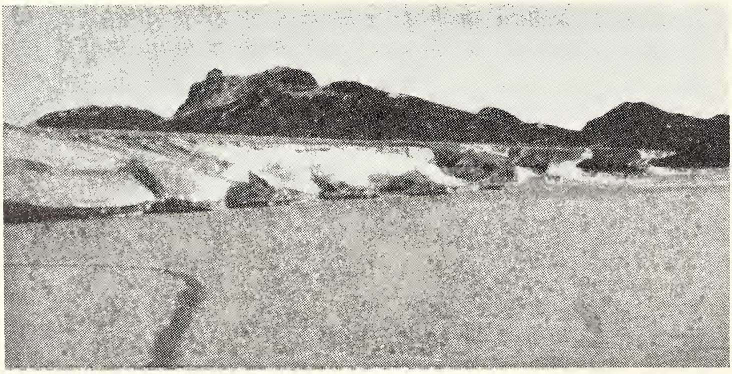

Fig. 30. The border of Hagafellsjökull (eastward from the place seen in Fig. 9, p. 23) facing the lake Hagavatn. During high water stage (in summer) the warm surface water has melted and abraded a deep trough into the ice. Overhanging ice is partly broken into floating blocks. Photo Helgi Jónasson, 1951.

The border lakes dammed up by frontal moraines and the proximal sides of sandurs — termed »*lón*» by Icelanders — are situatedat a lower level than ice-dammed lakes and they do not hold as great an amount of water at any one time. In 1952 the lón of Hoffelsjökull was 23 m and that of Falljökull 45 m deep (Hjulström 1953 p. 189). The marginal lakes, however, form sedimentation basins, from which new streams carry the meltwater out to sea. The majority of these accumulations of water are missing from the topographic map of Iceland, for at the beginning of the present century, when the map was drawn up, the glacier outlets generally extended as far as the end moraines and their water ran off into streams that flowed without interruption out to sea. It is mainly in recent decades that behind the end moraines has there been exposed a low-lying trough, which continues on underneath the ice. Water fills the most depressed parts of this zone up to the level determined by the height of the sill of the river starting at the lón. This level is naturally situated at different heights in different basins. It also varies in the same basin at different times, inasmuch as the river rapidly erodes its sill and as the water opens up for itself new outlets at lower levels than before. The waterlevel sinks down by degrees, therefore, from the time the lake forms, but simultaneously the lake spreads out toward the glacier. Following the melting margin of the glacier, it moves steadily on and fills the area being exposed from under the ice. The water of the lake likewise submerges the lower border of the glacier or penetrates into crevasses and underneath the ice, thereby promoting the melting process. The ice at the waterline melts fastest, and there it simultaneously forms a sedimentation limit, for beneath it the ice surface begins to be covered by drift, which retards the melting. Earlier the sedimentation limit was, of course, situated higher than at the subsequent, lower stages of the lón. The basal ice buried under sediments disappears slowly by melting, often only after the area has already been drained of water and become dry land. Also the outwash deposits accumulated on the ice turn into marginal lakes. First steeply sloping hollows (Fig. 31) appear on the gravel bed. There is water at the bottom of them and gravel covered ice can be seen in their sides. Gradually these hollows expand, for the stagnant water in them warms up and wastes away the ice, thereby undermining the gravel bed. The earlier gravel plain thereupon is transformed into a group of lakes dotted with drift-covered isles of ice (Fig. 32). Gradually these melt and the lakes merge to form one large body of water. The ablation moraines sink lower in the same fashion, with the ice wasting away inside and water frequently filling up the depressions. The relation of the ground to the level of the water and to the height of the point of discharge of the lón determines whether the ablation moraine in process of formation rises above the water surface or is submerged.

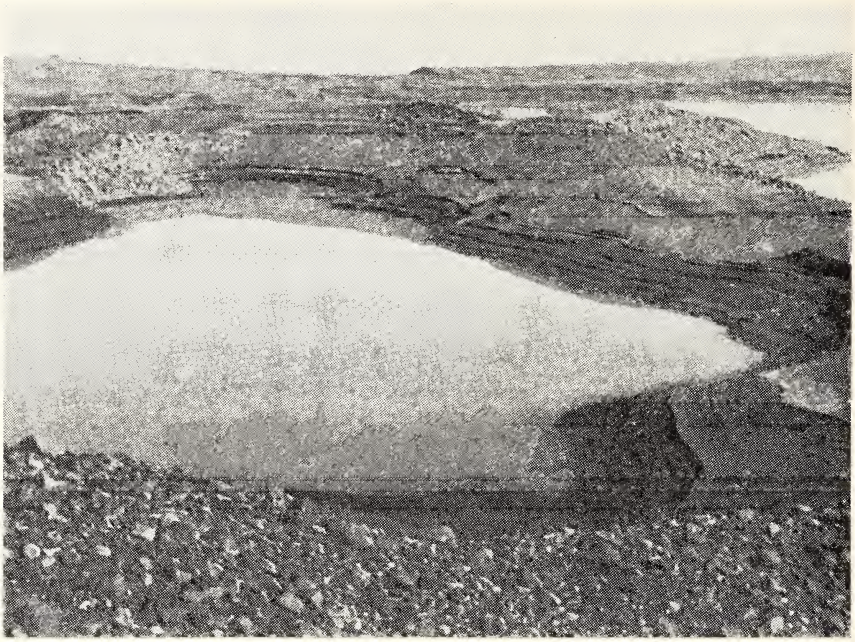

Fig. 31. A kettle hole in outwash deposits on Breidamerkursandur. Demarcation lines along the slopes indicate higher water levels. The diameter of the hole is about 20 m and its visible walls are 2.5 m high.

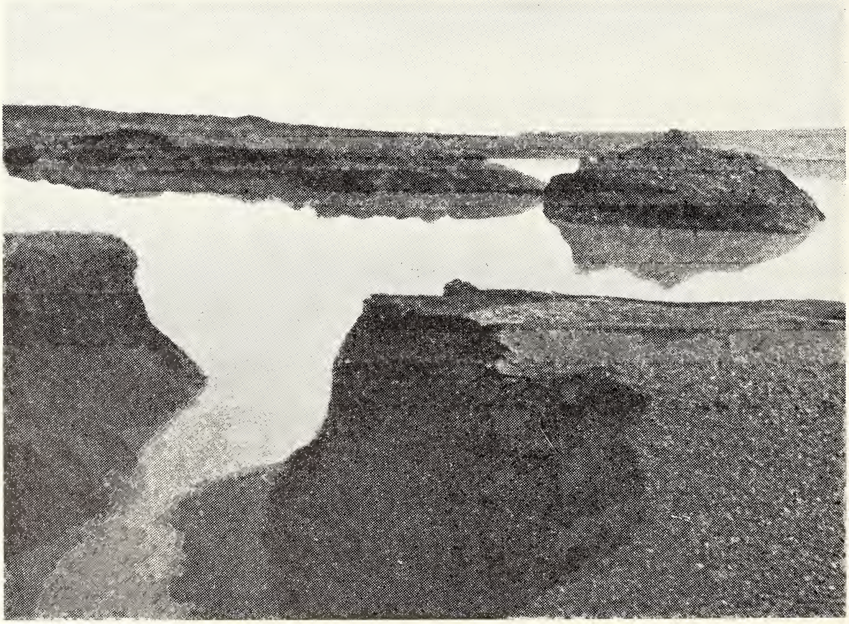

Fig. 32. A marginal lake at Breidamerkurjökull (near the hole presented in Fig. 31). The dark islands in it are drift-covered dead ice remnants. Dead ice underlies outwash gravel around the lake and becomes visible in the lowermost parts of the steep banks.

The waters in the radial crevasses of a glacier margin are connected with the marginal lakes, for the surfaces of both are situated on the same level (comp. p. 41). The water stagnating in the crevasses warms up and begins to melt the lower parts of the crevasse walls. Under such conditions a crevasse widens out like a cavern at the base. As the melting continues, the water undermines the ice and eventually joins the crevasses to each other, forming corridors and caves at the base of glacier. Thereby the lake expands underneath the ice and effectively promotes further melting.

A marginal lake is, in the view of the present author, an accumulation of water subject to continuous changes. The height of its waterline is determined by the height of the sill at the point of discharge, and its dimensions depend on the size of the low-lying area behind the sill. The quality of the glacier margin, on its part, determines to what extent water accumulates within and underneath the ice as well as in the crevasses and to what extent it is dammed up in the area from which the glacier ice has retreated. From the standpoint of glacier drainage the chief significance of the accumulations of water probably lies in the fact that they represent reception centers in which the flow of water from a glacier halts for a time before continuing its course towards the sea.

During glacier outbursts the marginal lakes lose their importance for under such circumstances so much water is discharged from the ice that the flow continues uninterruptedly across the lake basins, wearing down the sills damming the lóns.

BRAIDED RIVERS

The obstruction damming a lón, whether it be a frontal moraine or the proximal side of a sandur, is usually highest in front of the center line of an ice tongue and grades down toward either flank. As a result, waters accumulating behind an obstruction flow laterally between it and the ice margin until the lowest point in the obstruction is reached. The water is discharged over this point and continues on its seaward course. The starting points of the majority of the streams originating in the southern outlets of Vatnajökull are situated along the side of the ice tongues. The Jökulsá river, in front of Breidamerkurjökull, is an exception, for it intersects the end moraine in the center. In the view of the present author, this river has been active since the end moraine was formed, thereby preventing the moraine from obstructing its flow. The western branch of the Kviá river, starting from Kviárjökull, flows likewise through a deep channel in the frontal moraine. Several of the streams deriving from Skeidarárjökull also cut through moraines along similar channels. The normal volume of water in these streams is so slight that it could hardly suffice to wear out the present channels, but they must have been created during glacier outbursts. Such frequently recurring heavy floods in the frontal terrain of Skeidarárjökull pass over the marginal moraine at many points and wear down their sills so deep in the moraine that the existing rivers can discharge over them. In front of this glacier, therefore, as large lóns as those e. g. in front of Breidamerkurjökull have not been able to form.

The stream starting from a marginal lake discharges either over a rocky sill, thereby forming an erosive waterfall, or along a channel cut by it through glacial drift, whereupon the current is not so strong — in the latter case the river forms a swift stream rather than true rapids.

The water flowing over a rock sill erodes soft rocks of Iceland so rapidly that a canyon is produced relatively fast. Ultimately the river forms rapids at the bottom of such a steep-walled canyon. If the lake produced along the margin of a glacier empties out so fast that the sill has not enough time to flatten, a »dead waterfall» is left in the rock. Such relics are met with e. g. along the western side of Hagavatn, indicating the points where the lake had discharged its waters in the past (Fig. 4 p. 12).

Along the southern margin of Vatnajökull rock sills are rare, for the streams there flow out of the lóns into the sea through channels worn out of glacial drift. At only a few places near the glacier margin are there rock sills. In the summer of 1949 lake that had formed by the western flank of Hoffelsjökull emptied over such a sill into another, situated lower down (Fig. 29 p. 90). The flow had evidently started quite a short time before, when the sill had become exposed out of the ice, for it was only slightly eroded, although the lake had sunk down several meters, bringing into view the glacier foot extending out into the water.

A stream flowing over a sill formed of glacial drift usually cuts a vertical-walled channel for itself, the base of which is several meters below the surrounding ground level. The water is discharged through such a steeply sloping channel in a rapid flow, eroding the sill. The erosive (and transport) action of a river actually begins at this narrow point, for the water in a lón is for practical purposes stagnant. Usually the water accumulating in a lón area is discharged in a single stream, the water volume of which includes the entire run-off of one outlet glacier. On the basis of measurements carried out already by Ussing (1903) and subsequently by Thorarinsson (1939 b) and Hjulström (1953), the volume of water in rivers varies greatly. There is least water in the rivers in the winter and most in the summer. The largest rivers, Jökulsá and Skeidará (the chief river emanating from Skeidarárjökull), carry 300—400 m^3 of water a second in July (Thorarinsson op. cit. p. 211), but an even greater volume of water is estimated by Thorarinsson to have been drained out of Hoffelsjökull in the period July

29—30, 1937, namely 400—500 m^3/sec. This figure surpasses the maximum run-off measured by Hjulström's expedition in 1951 — 240 m^3/sec —, which was caused by a subglacial outburst of Gjánupsvatn into the Austurfljót. The level of the lake sank 25 m on this occasion (Hjulström op. cit. pp. 171 and 173).

The run-off of rivers also varies greatly during the different times of the day. There is most water in the evening and least in the period just before dawn (Thorarinsson 1939 c p. 209). These variations are, however, slight, compared to those at the different seasons of the year.

The variations in the volume of water reflect, at the same time, changes in rate of flow, which for its part affects the transport capacity of the river. When the capacity diminishes, the river unloads the mineral matter carried by it, deposits sediments on the bottom and begins to meander. Its course becomes winding and streams branch out. As the current quickens, it transports the sediments and continually changes its course. Both the main stream and its countless branches wind serpentlike toward the sea down the sloping plain. Hjulström has emphasized the lateral erosion action of glacial streams as creators of the strand flat (1954 b). Gradually the gradient of the river diminishes, the erosive power of the flow decreases and sedimentation becomes the chief operation. The branch streams broaden out and flow together to form shallow lagoons, separated from the ocean by a long offshore bar. At only a few points along the length of the bar is it broken, and through the openings the water emanating from the glaciers flows out to sea.

In the submerged coastal plain beyond the river mouths, there are depressions resembling submarine valleys radiating outward from the shore. They expand and deepen out to sea (Fig. 2 p. 9). The valleys are clearly visible at depth contours of 150 and 200 m, whereas the depth curves of 100 and 50 m do not yet reveal their existence, but run parallel to the shoreline. Thorarinsson (1937 b p. 166) considers these valleys the submarine extensions of fjords filled with glacifluvial deposits, expressing at the same time the thought that they could have been eroded by glaciers during times past when the sea level was 100 or 200 m lower than at present. The depth curves drawn at intervals of 50 meters do not make it possible to judge whether the valleys are the result of glacier erosion or river erosion; however, the depressions are too extensive to represent tectonic valleys. They greatly resemble the submerged extensions of fjords along the steep eastern, norhern and western shores of Iceland, but they differ from them in the respect that the depressions in the fjords are visible even in the 50 m depth curve. The difference is caused, as the present author understands it, by the fact that the valleys along the southern coast have been filled in the shallow coastal zone with sediment deposited by the currently existing rivers. Taking into consideration the present volume of water in the rivers and the rate of flow of the water emptying into the sea through the shore rampart, it appears probable that under prevailing conditions the flow weakens already in the coastal zone to the extent that sedimentation begins. The erosion of the submerged valleys must, accordingly, have taken place earlier, when the basal level of erosive action was situated considerably lower than at present. The extension of the valleys to a depth of at least 200 m does not, however, necessarily mean that the level of the ocean was so low at the time the valleys originated, for the volcanism of Iceland could have brought about vertical changes also in the situation of the coastal plain.

GLACIFLUVIAL DEPOSITS

The water drained out of glacier washes and sorts glacial drift released from the ice at as early a stage as the flow of meltwater over the glacier surface (see p. 36) as well as in subsurface streams and rivers outside the ice mass. During such treatment the glacial drift changes and begins to differ in both composition and structure from the till directly freed from the ice. Whereas lack of sorting characterizes the latter material, washed drift is sorted. Certain grain sizes have decreased in amount, while the remaining components have correspondingly been enriched. The structure of the strata formed of sorted material varies vertically in such a way that layers of different composition alternate. The sorted material, in other words, is also stratified. The designation *stratified drift* is, to be sure, often used to represent the sorted components of glacial drift. Stratified drift is produced by other agencies besides fluvial action (comp. Flint 1947 p. 103); hence stratified drift is broader than glacifluvial matter as a term, including e. g. eolian sediments. Inasmuch as the action of the water discharged by glacier decisively affects the differentation of Icelandic glacial drift, the accumulations resulting as an outcome of this action are herein termed *glacifluvial deposits.* The term thereby also includes the sediments of ice-dammed lakes and lóns. Only when it has been desired to emphasize the structure of the material has the designation *washed drift* been used; and »stratified drift» has been employed only as a stratigraphic term to indicate that the deposit comprises different strata.

As demonstrated in the previous chapter, the flow of water discharged from glaciers takes place nowadays in different stages, each of which affects in its own way the treatment of glacial drift. Consequently, in dealing with glacifluvial deposits, this study is observing a corresponding division, and deliberately deviating from the general grouping of these deposits. First, we shall take up for examination the accumulations forming in water drained off glaciers. Then we shall turn our attention to the sediments in the marginal lakes. And finally we shall consider the deposits of the rivers starting from these lakes.

DEPOSITS AT THE ICE MARGIN

Glacifluvial accumulations form already both on the surface of the ice and also evidently subsurficially. Sorted drift accumulates on the surface of the glacier as a result of washing and the action of surface streams. Subsurface deposits, on the other hand, do not come into sight except at the edge of the glacier.

Surface washing was described previously, on p. 36. The amount of water taking part in the surface washing process at any one time is so slight that its flow does not transport other than the finest grains. As a consequence of the irregular melting of the glacier surface and the washing of the finegrained material, the stones and even boulders fall into labile positions and roll down into depressions, where they become mixed up with the sorted drift. The sediment forming on the ice often therefore contains scattered stones. Furthermore, the structure of the sediment becomes disturbed upon the ice's melting underneath.

Surface drift rolls and slides also into the channels of surface streams. Near the ice margin, where there is more surface drift than higher up on the glacier, there forms here and there at the bottom of channels a unfied row of stones, from which the water washes the finer material away. If the current is strong, the water also transport the sand away from the stream bed and leaves only the stones in place. If, again, the current is gentler, the spaces between the stones fill with sand. A line of stones of this kind at the bottom of a V-valley was met with in the summer of 1949 at Breidamerkurjökull (Fig. 15 p. 38). Its stones were rounded only at the edges. The stones were 2—10 cm in diameter. The flow was so slow and the volume of water so slight that the stones remained stationary, and the water eroded them only through the agency of the sand running downstream. The line of stones extended up to the margin of the glacier, where the stream was discharged down a waterfall 1.5 m high onto ground moraine situated in front of the ice mass. On the surface of the moraine there could be seen, as an extension of the channel of the stream, a miniature esker (Fig. 24, Plate XIV), the breadth of which was about 50 cm and height 10—15 cm. Its material was composed of washed stones and sorted sand. The little esker could be followed for about 15 m, after which it sank lower and merged with the moraine. The stones of the ridge were unoriented, which evidently resulted from the fact that they had fallen off the ice margin on top of the moraine. At the bottom of the waterfall the material still underwent washing and the stones only then attained the position in which they were left on the little esker.

The miniature eskers were rare, however, for ordinarily in the channels of streams flowing over the glacier surface there was so little drift that no accumulations whatsoever formed in front of the ice margin. For example, the streams flowing off the margin of Hoffelsjökull as little waterfalls washed only the surface of the ground moraine. As this surface sloped toward the glacier, the water continued its flow under the projection formed by the ice margin and passed out of sight. From under the waterfall there formed a row of stones on the surface of the moraine leading under the projection. It will be deformed, however, as soon as the projecting ice collapses. Since the volume of water in the surface streams was extremely small in the summer of 1949 and their transport capacity was limited, the accumulations were more in the nature of curiosities than glacifluvial deposits, properly speaking. They indicate, however, that under certain conditions glacifluvial deposits may form, as an outcome of the action of surface streams, both on top of the ice and in the frontal terrain of a glacier.

Subsurface waters constitute the chief part of the flow in the drainage of Icelandic glaciers (p. 81). The water ran out of the glacier tunnels so gently and lacking in hydrostatic pressure that its capacity to transport material was exceedingly small. The water was, to be sure, turbid and contained an abundance of fine sand and silt, but the flow could not be observed to carry along stones from underneath the ice even along the bottom. If subsurface water did transport glacial drift, stones, gravel and sand, this action terminated before the water was discharged from the ice margin. It is possible that glacifluvial material was deposited in crevasses and tunnels (comp. Todtmann 1951, Hoppe 1953 p. 263, Ives and King 1955 p. 480), through which the water runs. The filling up of tunnels with glacifluvial material is not, however, a general phenomenon in the glaciers of Iceland, as indicated by the fact that the glacier tunnels of Morsárjökull and Hagafellsjökull were empty of glacifluvial material in the summer of 1949 (p. 82).

Nor did stones and gravel appear in the water forming springs beyond the visible margin of the glacier. At the »eye» of such a fountain, however, the flow was so strong that it lifted, e. g. in front of Breidamerkurjökull, stones measuring 2—5 cm in diameter above the level of the stream and kept under 10 cm stones constantly moving to and fro. These stones originated, however, in the deposit through which the water was discharged, for no glacifluvial material could be detected carried by the water into the fountain. The water of the fountains spouting out of the ice was still clearer.

The observations described in the foregoing lead the present author to the conclusion that under normal conditions, such as prevailed in the summer of 1949, the water draining from the glaciers of Iceland transports only fine-grained drift.

DEPOSITS IN MARGINAL LAKES

Marginal lakes and other accumulations of water at ice margins receive most of the glacier drainage. As the rate of flow diminishes or ceases altogether in the basins, deposition of the material transported takes place. Sediments lie submerged until the water level sinks and the bottoms first come into sight after the lakes have been drained off.

Ice-dammed lakes at the eastern side of Hoffelsjökull had become emptied just before the author visited the place (July 14, 1949) and their bottoms were still bare. The author could study the bottoms of the lakes of Efstafellsvatn and Gjánupsvatn. At the bottom of the former there flowed a river emanating from the mountains toward the edge of the glacier along a steep-walled channel, which had been worn down into the bottom sediments. These deposits contained alternating beds of stones, gravel and sand, cutting each other at different angles (Fig. 29, Plate XV). At one point the whole strata was bent, saddle-like, and at other points distinct cross-bedding could be seen in the sandlayers. All the beds contained some scattered stones, a few of which were rounded and the others subangular. The most flattened of the stones had an orientation parallel to the slope of the strata. From the inclination of the layers and the cross-bedding, it could be seen that the ice-lake deposits had been carried to the basin by the river from the mountains. It is probable that the irregularities in stratification are due to the proximity and instability of the ice margin. For the same reason the lake bottom is uneven, containing gently sloping knolls of gravel and sand with depressions in between. In the present author's opinion, they originated during outbursts of the lake, which cause the ice margin to collaps and scatter an abundance of floating blocks of ice. The topmost part of the deposit contained coarser fractions, in general, than the lowermost parts. At the bottom of the section there were sandy deposits (Fig. 33, Curve 13, p. 100).

The bottom deposits cover the foot of the hill, too; but when the slope turns steeper, the sediment mantle disappears, until a narrow gravel bench marks the water level of the lake. Seen from a distance, the benches form rather plain horizontal lines, but closer up they become scarcely distinguishable (Fig. 30, Plate XV). Likewise insignificant, though perceptible from a longer distance, benches occur along the shore lines of other ice-dammed lakes as well. Upon comparing these with the ancient shores of Finland and Sweden, they are weaker and were the land surface not so bare, it would be quite difficult to observe them at all.

The bottom of lake Gjánupsvatn resembled the former in that the surface was there, too, uneven, though sections were lacking. In some places a thin layer of sorted sand and gravel was to be found on the surface, which generally consisted of till, with small angular rock fragments. Kettles described on p. 56 had been imprinted on the soft bottom. Many of them still had melting ice blocks, but in addition to these fresh ones, ice-free kettles of older origin, with gentle slopes, dotted the bottom.

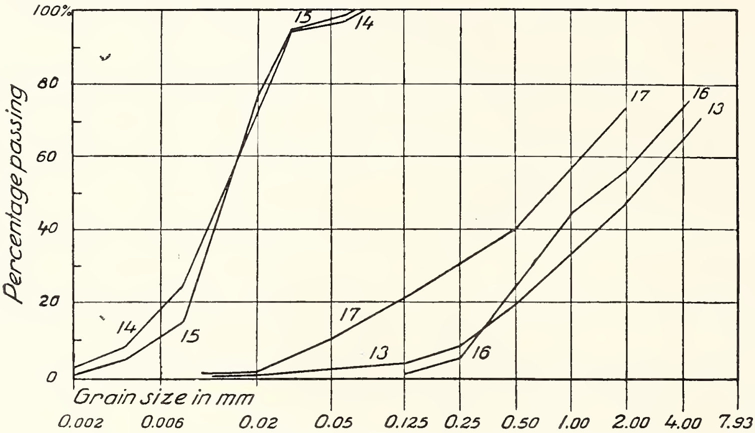

Fig. 33. Grain size distribution in glacifluvial and glacilacustrine deposits. Curve No. 13 ice lake deposits in Efstafellsvatn, No. 14 varved ice lake sediment, east of Hagavatn, Langjökull, No. 15 ice lake deposit from the same locality, No. 16 glacifluvial deposit from an erosion esker, Morsárjökull, and No. 17 drift from the sandur plain in front of Hoffelsjökull.

The third bottom of an ice-dammed lake was laid bare in the valley on the east side of lake Hagavatn at Langjökull. The ice barrier damming the lake had melted some thirty years before, causing the lake to drain off. Shore benches were still visible, showing that the ancient waterline had been situated about 45 m above the valley floor. Along this valley there flowed a river fed with meltwater from the glacier. In some places it had worn its channel into the lake bottom, while in other places it had spread its own deposits on the lake sediments. A flat-topped gravel knoll on the western side of the river displayed the following stratigraphy: the top was composed of stratified gravel and sand to a depth of fifty centimeters. It was followed by cross-bedded coarse sand (60 cm), which became finer downward and contained thin laminae of silt. Underlying this there followed a layer of varved sediment 30 cm thick. The varves were regular and varied between one and two centimeters in thickness. Each varve had a lighter (coarser) and a darker (finer) part consisting mainly of silt (Fig. 33, p. 100). As the present author understands it, they were true annual varves, resembling the diatact varves found in Finland (Sauramo 1923). The deposition of such sediment in a marginal lake can be conceived of by assuming that the lake was covered with ice in winter (as Hagavatn is nowadays).

As the meltwater activity diminished at the same time, the silt suspended in water could deposit and form the fine-grained winter parts of the varves. In this respect the sediment differs from the ice-lake clays of Denmark, which have much thicker and more irregular varves, as described by Hansen (1940). The same difference was observed also by the author in the summer of 1951, when he had an opportunity to see some Danish ice-lake clays under the guidance of Dr. Sigurd Hansen.

Beneath the varved clay there was a layer of silt 25 cm thick, the bottom part of which was stratified. This stratification was caused by thin laminae of silt and fine sand. A sharp boundary separated this sediment from the underlying gravel bed. Judging by the stratigraphy, the ice-lake has had a short passive stage, during which varved sediment could deposit. Similar »very beautiful warved clay» appeared at the bottom of Hagavatn after its drainage in 1939 (Thorarinsson 1939 p. 238). From the marginal lake of Langvatn (west of Hagavatn) Wright has described and photographed »the typical varve or banded clays deposited in glacial lakes, presumably due to the different rates of sedimentation in summer and winter» (1935 p. 225). In addition to these, varved clay has been found in southern Iceland, where short profiles have been measured and compared to the Swedish time scale by Ebba Hult De Geer (1928) and J. Áskelsson (1930).

From the afore-mentioned varved sediment and from the silt layer beneath it pollen and diatom analyses have been made. Here is the result of pollen analysis:

	Varved sediment		Silt	
	sp.	%	sp.	%
Betula	7	17.5	14	30.4
Pinus	1	2.5	1	2.2
Salix	11	27.5	—	—
Ericales	—	—	1	2.2
Cyperaceae	10	25.0	12	26.1
Graminae	6	15.0	10	21.7
Caryophyllaceae	—	—	2	4.4
Varia	5	12.5	6	13.0
	40	100.0	46	100.0

In addition, 5 *Lycopodiaceae* and 4 *Polypodiaceae* spores were met with in the silt sample.

The samples contained the following diatoms:

	Varved sediment	Silt
Achnanthes sp.	—	2
Cymatopleura solea	—	1
Cymbella sp.	1	—
Epithemia intermedia	1	—
» *zebra var. saxonica*	1	—
Eunotia praerupta	3	1
» *sp.*	1	1
Fragilaria sp.	6	4
Gomphonema acuminatum	—	1
Melosira ambigua	—	3
» *italica*	—	1
Navicula sp.	—	2
Pinnularia borealis	3	4
Rhopalodia gibba	1	—
Synedra sp. fragm.	2	—

The diatoms in both samples represent scarce shallow and fresh water flora. It contains mainly the same diatoms that exist at present at Vatnajökull and Langjökull (Mölder 1951).

The marginal lakes, or lóns as they are called in Iceland, have evolved between the ice border and end moraines. In these lakes the main part of the material carried by meltwaters from the glacier is unloaded. The bottom of the lake is, however, so soft, long after the drainage of the lake, that it is difficult to study its deposits. In front of Skaftafellsjökull a lake bottom was laid bare, forming an undulating area between the visible ice border and the end moraine zone. This area differed from the ablation moraine in the respect that the topmost layer in the soil profile consisted of stratified sand. Stratification was caused by the altering of sand and silt layers. The sediment resembled the Danish ice-lake clays (Hansen 1940). Some separate stones were imbedded in the sediment and in some places unsorted material could be observed. Probably this had slid down into the lake from inclined ice surfaces, as such sliding processes were seen in action. The frontal part of Skaftafellsjökull was buried in sediments up to the earlier water level and the gradient of the ice surface was so slight that no sliding of drift happened there. Disturbances appeared only farther away from the visible ice margin in the zone where the buried ice had begun to melt. Here steep-walled holes occurred in the sediment plane, resembling those in ablation moraines (p. 54). Compact, dark dead ice was found in the lower part of the walls, but now it was covered with stratified drift. From the steep walls drift slid and slumped down into the holes, thereby bringing new ice into sight. At the same time water collected in the holes; and it melted the bordering ice so that the holes were widened until they joined together. The lake bottom, which was primarily rather flat, thus turned uneven and its base level was lowered. The final result of this development must be a formation whose topography is irregular and material composed of ice-lake deposits. After drying, these sediments offer the wind erosion a very suitable surface.

Similar lón sediments were visible along the shores of the marginal lakes in front of Hoffelsjökull, Breidamerkurjökull, Kviárjökull, Morsárjökull and Skeidarárjökull. The lowest parts of the lake basins were still submerged and deposition took place there. Because the ice-lake deposits cover also the emerged land, the water levels must have reached higher before the present stage. The draining off of the marginal lake in front of Svinafellsjökull had laid bare long and narrow land tongues and islands covered with lake deposits. The whole landscape thus resembled dead ice topography. The development at Morsárjökull, as seen in 1949, must bring about a similar landscape, when the marginal lake flooding over the thinned ice margin has been lowered (Fig. 25, Plate XIII).

The appearance of marginal lakes is a result of rapid shrinkage of the glacier, which happened during recent decades. When the ice margin extended to the frontal moraines, the topographical conditions prevented water from collecting at the margins; and the water ran without any obstruction to the sea. The significance of the marginal lakes in Iceland has been stressed in recent years by Hjulström (1953, 1954 a and b). It is evident that the marginal lakes have a great effect on the melting process of the ice terminus and on the deposition of glacifluvial deposits. Only dissolved material continues its way through the marginal lake without any sedimentation.

DEPOSITS IN BRAIDED RIVERS

Rivers emerging from the southern outlets of Vatnajökull have filled with their sediments valleys extending from the interior to the sea. These sediment fillings have been known as sandurs (the Icelandic name in the singular is *sandur*, the plural form being *sandar*). Many scientists have made observations on them as they have crossed the vast sandur areas along the southern margin of Vatnajökull. The best known sandur, the Hoffelssandur (Thorarinsson 1939 b, Hjulström 1953, 1954 a and b), has been built up by the Austurfljót and Vesturfljót, which drain Hoffelsjökull. According to both investigators, the gravel plain is a result of the erosion, transport and deposition activity of these rivers.

As early as 1881 Helland gathered water samples from Icelandic rivers and investigated their silt content. Like other glacifluvial streams, they were heavily loaded or braided. A water sample taken from Nupsvötn contained 570 mg silt per litre, and that from another river emerging from Skeidarárjökull contained 1 509 mg/litre. A water sample from the Jökulsá at Breidamerkurjökull had a silt content of 1 876 mg per litre (Helland 1882). Thorarinsson estimated that the Austurfljót carried in the summer months of 1939 on the average 1 070 mg silt per litre. The high average of ignition loss (3.2 %) was, according to Thorarinsson, caused by organogenic matter (1939 b p. 212).

After the forming of the marginal lakes, the transport of material has evidently diminished. The rivers beginning from these lakes cut their channels in end moraines and gravel beds eroding them. In the upper course of the river profile the gradient is so great that erosion prevails, while in the lower course deposition takes place. The transport capacity varies, greatly in time and in different parts of the stream, causing an alternation in deposition and erosion even in the same place. For that reason the rivers are continuously changing their channels, meandering and winding like pendels along their beds.

The surface of the outwash plain is filled with dry stream beds and delta islands between them. The top-planes of these islands dip downstream. So do the stream channels, too, and they are so inclined that it is easy to determine the direction of flow. Old, low-walled channels cross diagonally even the top-planes of the delta islands, which usually are triangular in shape. The sharp angle is directed upstream while the base faces downstream. The flanks slope rather steeply toward the bordering flat stream beds, which commonly lie at different levels on both sides. The erosion remnants have not always the shape of a triangle. Sometimes the stream erodes the plain from one side only, thus cutting a steep bank into it. Such banks separate two gravel plains, e. g. in front of Falljökull; and a similar bank has formed alongside the river Jökulsá at Breidamerkursandur.

The stream can also wear the plane from two sides so heavily that the dlane becomes narrower and narrower, until the erosion remnant has the form of an esker. Its slopes are at first very steep, but later they take an angle of repose. The original plane now forms a very narrow flat top between two steep-sloping sides, and it has a gentle inclination downstream. In many places there are still weak channels on the flat top intersecting it at sharp angles. If the erosion continues so long that it wears away the plain totally, the erosional remnant gets the shape of a sharp crest or esker ridge. Owing to the course of eroding rivers the same »erosion esker» is likely to contain sharp crests and flat-topped narrow ridges joining together.

Erosion eskers of this kind occurred in 1949 in front of Morsárjökull and Langjökull. Their length was not more than 150, their height varied from 5 to 10 m and the basal width from 20 to 80 m. The base levels or bottoms of the slopes were at different heights and very often a channel formed by the eroding river separated the erosion esker from the lower-lying younger outwash plane. The erosion eskers were commonly directed outward from the glacier. An exception was made by a ridge in front of Morsárjökull. It started from the southeastern border of the valley and crossed the valley train diagonally downstream until the present river destroyed it. Judging from the erosion channels, this river had earlier emerged from the southeastern edge of the glacier and streamed laterally between the frontal moraine and the glacier margin. During this stage of development the river had worn the earlier delta plane and formed the erosion esker. At a later stage a marginal lake developed at the glacier border and the river now begins from the northwestern end of it. The water level is some meters lower than the top of the esker ridge, showing that the basal level of sedimentation has considerably lowered. This is quite a common feature in the proximal parts of Icelandic sandurs. In the author's opinion this part of the outwash plains was deposited under conditions when streams were stronger and carried more material than they do at present.

It was mentioned earlier (p. 103) that the rivers carry from the marginal lakes suspended and dissolved matter only. The stream very soon grows so strong, however, that it begins to erode the bed already from the sill, cutting its channel into the drift. Especially wet moraines, as they are after being released from the glacier (Fig. 31, Plate XVI), and river deposits are very suitable objects for stream erosion. The stream spreads the material and carries the finest fractions away, leaving the coarsest particles *in situ* or carrying them short distances only. As the strength of the current varies greatly, the newborne deposits become stratified and very often a certain lens-structure appears in them. On the other hand, the deposits are weakly sorted and their grading resembles that of till (Fig. 33, curve 17, p. 100).

In places where the current has been violent, as e. g. in the present discharge of Hagavatn during its formation in 1939, the stream bed is filled with blocks. These have a very regular orientation: their long axes are turned either in the direction of flow or then lie perpendicular to it. The flat topsides always turn upwards to the direction of flow. This kind of orientation is not restricted to the surface; then very often it becomes visible also in cuttings (Fig. 34 p. 106). The coarse block fields are concentrated in the proximal parts of the outwash plains, i. e. in front of end moraines. Behind them block fields were seen on the old deposits of glacier outbursts.

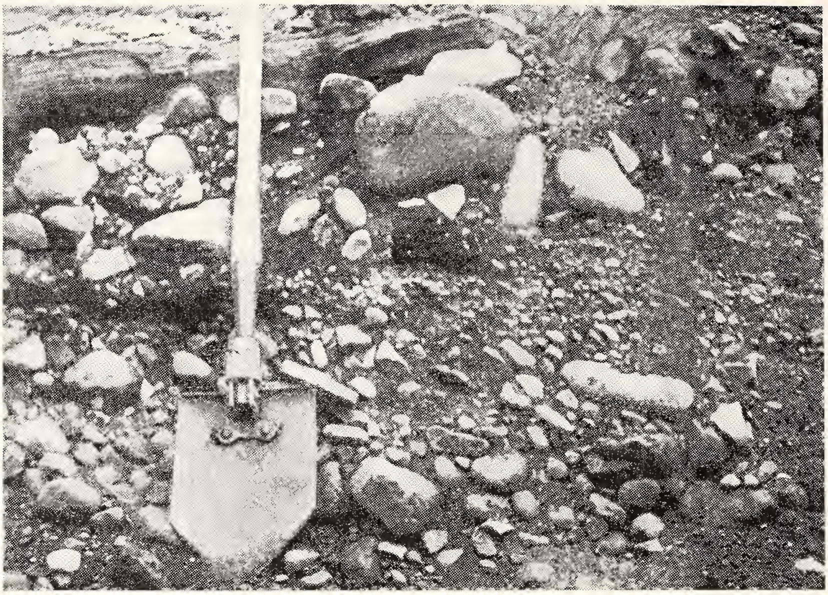

Fig. 34. Structure in a sandur in front of Heinabergsjökull. The stream has flowed (from right to left) to the direction indicated by the upturned ends of most of the stones.

In front of the eastern margin of Breidamerkurjökull a block field was met with in the proximal side of the end moraine arc. Its flat top was some 3—4 m higher than the water surface in the present lateral river. In this block field some peat chunks were lying among the boulders (Fig. 32, Plate XVI). They consisted of *Carex-Bryales* peat with the following pollen content:

	sp.	%
Betula	23	33.3
Pinus	1	1.4
Cyperaceae	35	50.7
Myriophyllum alterniflorum	1	1.4
Salix	1	1.4
Varia	8	11.6
	69	99.8 %
(*Polypodiaceae*	1)	

Farther away the size of the carried material gets smaller, after which gravel and sand begin to prevail in the surface material. The stones at the bottom of stream beds have settled so that their longitudinal axes are perpendicular to the stream. This comes out in the orientation analyses

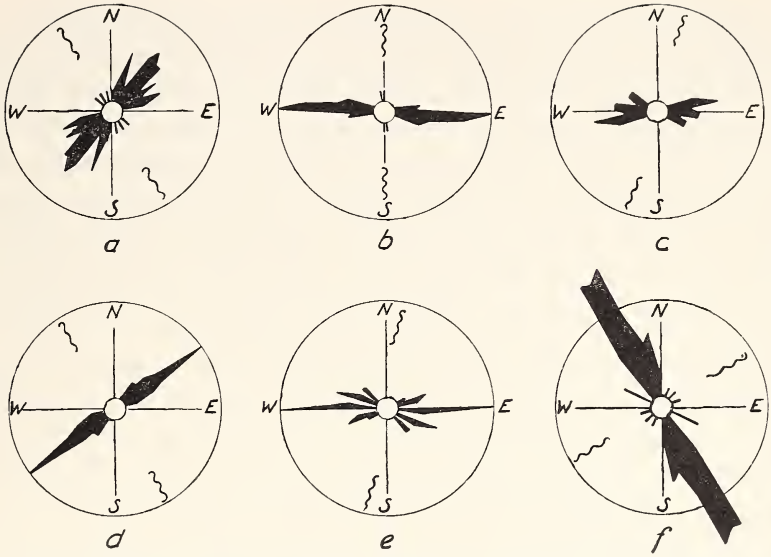

Fig. 35. The orientation of the sandur material: a. and b. on the surface of Hoffelssandur, c. in front of Heinabergsjökull, d. in a dry stream bed on Breidamerkursandur, e. in the same at Ignolshöfdi, S. of Öræfajökull, and f. in coarse outwash material at the lower end of the present gorge of Hagavatn. The direction of flow is indicated by a winding line in each diagram and it is determined from the trend of stream beds. The radius of the circle signifies an orientation of 20 %.

(Fig. 35) compared with the direction of flow marked in diagrams on the basis of the trend of the stream bed. Very often these dry beds cut the direction of the present rivers at sharp angles. In lateral stream beds the stone orientation had in some cases the same trend as the probable ice movement, though the former has, of course, nothing to do with the latter.

The sandur stones are in general rounded and they lack striations. Locally, however, there are plenty of angular stones and blocks on the sandur. They occur in such places where the material, deposited during the glacier outbursts forms, such elevated planes that the normal flow of the river has not been able to destroy them. The size of carried material indicates powerful stream action, while the shapes point to short transport also of short duration.

The angular material is mixed with the rounded stones also in places where the river erodes moraines in front of the glacier and spreads their till, washing it in the same. The coarse fractions are then carried such short distances that they preserve their shapes.

The third occurrence of subangular or angular material constitutes the fans of mountain rivers, which in many places adjoin outwash plains. These rivers carry, especially in spring, angular rock fragments from mountain slopes, and they have not time enough to become worn round before depositing.

As a whole the Icelandic sandurs are, in the author's opinion, supra-aquatic deltas or outwash plains with a downstream inclination, and their surfaces are crossed by countless stream beds. In cross profile the surface of a sandur is uneven and its central cone rises higher than the margins (Hjulström 1953 s. 179). This is plainly to be seen especially in the proximal part of the outwash plain, where the present rivers emanating from marginal lakes flow along the sides, eroding the lateral parts of the plain. The central cone joining the end moraine lies always higher than the present erosion level. The long profile shows the greatest gradient in the proximal part (op. cit. p. 178). Towards the distal part it diminishes so that the long profile is hyperbolic. According to Hjulström's measurements, the gradient of Hoffelssandur is in its proximal part 10 m per kilometer, while in its distal part the mean value is 0.5 m per kilometer (op. cit. p. 178).

The fact that the sandur plains reach over the present water level and contain in their proximal parts a very coarse material indicates that these parts have been formed as a result of a more powerful stream action than at present. During this former stage of development the glacier front remained in the end moraines. Since the glacier margin retreated from that position, the marginal lakes have diminished the glacifluvial action and hindered the outwash plains from expanding to the present ice margin. At the same time the share of carried material has diminished to such a degree that the present rivers mainly rework old deposits and transport their drift towards the sea. The finest particles are also now carried directly into the sea. Clay is lacking in the supra-aquatic part of the delta and its deposition takes place first under the sea level. This subaquatic part of the delta is separated from the supra-aquatic part by a sand bar which dams lagoons at the river mouth. They are connected through narrow sounds with the sea. The water in lagoons is seemingly stagnant but levelings show that their surfaces have a gentle seaward slope. Hjulström's map (1954) shows that the water level has a height of 10 m at the uppermost end of the lagoon, while the lowermost end lies almost at sea level. The water masses behind the bar are thus slowly moving towards the sea.

The sandur consequently consists of three parts: supra-aquatic outwash plain, offshore bar, and subaquatic delta. The proximal part of the sandur adjoins the end moraine belt, but some water-eroded channels connect the sandur to smaller outwash plains deposited at the ice margin and, in some instances, even on the ice.

ON THE DEFORMATION OF DRIFT DEPOSITS

Accumulations of glacial drift seldom preserve their original forms and structure. Commonly, perhaps always, the exogenetic processes rework the material outside the glacier to such a degree that the accumulations acquire new features. In other words, the accumulations of glacial drift are deformed after their deposition. In this process many agencies take part, among which the work of ice, water, and climatic conditions is decisive. Reference has already been made in many connections to these agencies; but to get an idea of their share in the deformation of glacial drift deposits, a review of these agencies is presented. It is, however, important to keep in mind that many agencies usually partake in the deformation process at the same time, though in varying degree.

The action of the ice is directly related to the behavior of the glacier. It appears in two ways: the deformation of glacial drift accumulations by the advance of the glacier, and the deformation of drift by the melting of the stagnant ice.

The margins of the Icelandic glaciers are in generally unstable and sensitive. The snouts of the outlets are pushed forward many times over the frontal terrain. In some places the ice margin advances over the drift accumulations, wearing their uppermost parts. As already shown, erosion does not reach very deep down into the drift, but the snout over-rides the deposits slightly. For that reason glacifluvial deposits are found under till cover in undisturbed conditions (p. 62). For the same reason the ground moraine has its primary structure and only the shape of the uppermost boulders indicate a later advance (p. 45). A glacier that has readvanced on push moraine loads it with a till layer without destroying its form (p. 64). The slight erosion of the glacier evidently depends on obstructed flow, which causes overthrusts in the frontal part of the glacier. The basal erosion grows stronger when the advancing ice is thicker. This erosion zone is not, however, under the edge of the glacier but somewhere more distant from the margin. If the oscillation does not reach so far that the deposits accumulated in the frontal area come into the erosion zone, these deposits retain their forms and their structure beneath the advancing ice front.

Glacial deposits accumulated on the ice undergo a much stronger deformation. Also the englacial accumulations deform as the ice melts under them. In many cases the ice cores are preserved so long in the superimposed drift that the deposits acquire their topography first after being released from the glacier. The originally smooth forms can thus change to irregular knob and kettle areas. This topography can evolve in both unsorted and sorted glacial drift, assuming that there has been buried ice in the drift. The present conditions at the Icelandic ice borders favor this development. Transversal moraines and clean ice sheets between them have caused a knob and kettle topography with a transversal orientation. Many end moraines (at Hoffelsjökull, Breidamerkurjökull, Skaftafellsjökull, Morsárjökull) have been deformed in this way. Kettle topography is produced also at the bottoms of ice dammed lakes, regardless of the kind of drift (p. 56).

The action of water in the deformation of glacial drift deposits begins already on the glacier, as the surface drift is washed by rain and meltwaters. The finest fractions are carried away, causing thereby an enrichment of coarser grain sizes in the remaining material. The variations in the quantity of water and in the velocity of flow are reflected in the deposition and erosion as well as in the stratigraphy of the drift. This process is repeated in the rivers emerging from the glacier as well in those which begin in the marginal lakes. Consequently, the deposits in both cases have a similar structure.

The deposits of both rivers correspond also morphologically as they form outwash plains. Because of the marginal lakes the transport of material diminishes and the erosion increases in importance. Especially in the proximal parts of the sandurs, where the gradient is considerable and the rivers flow along a few main channels, the river beds have been cut deep into old deposits, thereby eroding the surface of the outwash plain. Simultaneously, rivers erode their beds laterally as well (Hjulström 1953), forming steep banks. Between the river beds there form various erosion remnants, which at certain stages of development resemble eskers (p. 104); but the erosion is likely to continue so long that the original delta plane disappears and only the younger planes at a lower level remain. The surface of the outwash plain is thus deforming and lowering step by step as long as the action of the rivers continues. For that reason the surface does not always indicate the highest position of the water level, but it is a result of continuous river erosion. For the same reason the adjacent delta planes situated at a different height only signify various stages of development. The order of their age is in general such that the higher ones are older than the lower; but sudden floods, glacier outbursts, etc., can cause exceptions.

The river erosion is also directed towards the nonsorted drift at the glacier margin. This drift is so wet and soft that the river erodes it very rapidly. The ground moraine, after being released from the ice, loses its own topography and structure outside the glacier, changing to sorted material under the influence of river action. The end moraines are deformed in the same manner, too. The rivers cut their beds through the ridges during the floods, but during the normal flowage these beds stay dry. Only when the beds have been cut so deep into the moraine that the normal runoff can use them, the rivers flow directly from the glacier margin to the sandur, preventing the occurrence of a marginal lake.

The erosive action of the streams consequently works on the glacial drift quite effectively outside the glacier. It works up till into glacifluvial deposits, cuts end moraines, and wears down planes at different heights to the sandur surface. The deformation process continues as long as the rivers exist. At the present stage of development the proximal parts of outwash plains are lowering and their deposits are gradually carried into the sea.

The part played by climatic agencies in the deformation of drift is manysided and varies in different areas. Moisture, air temperature, and wind relations contribute to this process.

In moist areas, as e. g. in the southern part of Iceland, the drift being released from the glacier remains water-soaked for long periods. Along the slopes it flows slowly down. Flow earth occurs especially on the sheltered proximal sides of end moraines, which are little by little becoming smooth. The heavy precipitation also keeps the ground water level high near the glacier margin. As the drift, especially in glacifluvial deposits, is permeable, the marginal lakes drain subsurficially through end moraines and form pools on the sandur. Clean-water streams issue from these springs and they add to the drainage in braided rivers. Water fills the hollows in the dead ice, too, and in these kettles the fine-grained sediments settle down. As the walls of the kettles simultaneously slump until they achieve an angle of repose, coarser material is mixed with the fine-grained sediments. The originally steep-walled kettle holes become funnel-shaped.

Owing to the effective evaporation on the wind-sides, the soil surface very soon becomes dry. The distal sides of end moraines are much drier than the proximal sides. The drying up of the surface adds to the possibilities of wind erosion. The winds pick up mineral grains from the dry surfaces, thus forming gravelly and stony deflation planes, which cover the dried glacial drift deposits. The fine-grained material borne by the wind polishes the surfaces of blocks and stones, wearing off the glacial striations and causing instead corrasion surfaces. Most of the fine-grained material is picked up by the wind from the dried bottom sediments of marginal lakes and other accumulations of water. The »smoking» of sandurs is mainly caused by these deposits.

The southern outlets of Vatnajökull stretch so far down toward the sea coast that the winter frost does not deform the glacial drift deposits in the frontal area. Hjulström has found periglacial morphology on the Hoffelssandur only in a few places (1953 p. 180).

In the interior of the island the precipitation and the temperature are lower than along the southern coast. The ground water level lies in general deeper and the land surface is dry. The drift deposits freeze in winter and in spring their surfaces are water-soaked. Periglacial solifluction phenomena are common (Thorarinsson 1951). On the surface of ground moraine there occur frost polygons, stone rings, and other features commonly met with in arctic regions. Some marks of periglacial solifluction were seen in front of Langjökull at Hagavatn (p. 47). Because the winds are here dryer than in the coastal area and they blow over bare, open landscape, the wind erosion has a strong effect on the land surface. The fine-grained material disappears from the moraine surface and simultaneously frost wedging breaks blocks into pieces (Hawkes 1924). The slow deformation of glacial drift is thus under way in the central highland. Evidence of these processes is to be seen in the structure and material of old moraines, which are reworked, restratified and washed (Thoroddsen 1905—06).

COMPARISON WITH OTHER GLACIATED AREAS

Among the glaciers of the world today the largest ones in Iceland are in many respects in an intermediate position. In situation they are among the southernmost of the Arctic glaciers, and in size they rank among the large ones, if not among the very largest. In Klebelsberg's listing (1948 p. 209) Vatnajökull ranks fifth and Langjökull twelfth. Ahead of Vatnajökull come only the ice sheets of Antarctica and Greenland as well as the glaciers of West Spitzbergen and North East Land. After it come next the glaciers of Severnaya Semlya and southern Alaska (Malaspina). Also in form the glaciers of Iceland are in an intermediate position. They are glacier caps whose ice mass is so thin that the main body of the glacier reflects on its surface the irregularities at the base. Klebelsberg (1949 p. 582) applies to them the term *Plateaugletscher.* The outlets starting from the ice cap proper, which along the southern margin of Vatnajökull stretch out near the level of the ocean, resemble in many respects valley glaciers. The termini of the largest outlets, again, spread out upon reaching the coastal plain into expanded feet of the Piedmont-glacier type.

Climatically the glaciers of Iceland differ from the Arctic glaciers in that they belong to the temperate glaciers (Ahlmann 1933), whereas most others are either sub-polar or high polar. The temperature of the ice affects its physical properties, for the plasticity of the ice increases towards the pressure melting point. Consequently, temperate glaciers are faster moving. Since, in addition, the glaciers of Iceland receive more abundant snow than Arctic glaciers in general, they are active, although they melt rapidly (Ahlmann 1953). As a result of the heavy precipitation and, at the same time, warm climate, the ice wastes away, largely by turning to water. Rain and meltwater occur on the glaciers of Iceland in greater abundance than on other Arctic glaciers. In addition, local circumstances, mainly subglacial eruptions, cause exceptionally strong glacifluvial activity in certain parts of the Icelandic ice caps. Since the glaciers move fast, they crack up and the meltwaters run off the ice surface into the crevasses, being discharged eventually from underneath the margin.

It is natural that under the conditions described in the foregoing the formation, transport and accumulation of glacial drift should be different from what they are on glaciers of other types. Already Hobbs (1911) stressed the difference between mountain glaciers and inland ice sheets in describing the existing glaciers. »The determining factors in the case of mountain glaciers seem to be an upturning of the ice layers in this region and surface ablation or melting. In the case of inland-ice the important factor is the deposition upon the surface of snow borne by the wind from the interior of the mass» (op. cit. p. 289). Arctic ice caps are, according to Hobbs, in an intermediate position between these two principal types.

Glacier research of recent decades has moderated Hobbs' sharp boundary between inland-ice and mountain glaciers. Numerous examples have been presented from the Arctic towards proving that the margins of ice caps and ice sheets behave just about in the same way as those of Hobbs' mountain glaciers. This is perhaps because the present amelioration in the climate has begun to affect the glaciers and be reflected in the behavior of their marginal zones. According to Ahlmann (1953) the size of glaciers affects their reaction to climatic change in that the change is felt sooner on small glaciers. The valley glaciers of the Alps and Scandinavia began to melt already towards the end of last century, so that the descriptions of these glaciers from that period reveal the behavior patterns of a retreating ice margin. The Ameghino glacier in South Patagonia began to shrink in 1870—1880 (Nicols and Miller 1951). The culmination point of the Icelandic glaciers occurred in the 1890's, and the vast ice mass of Antarctica has not yet, according to data collected by the Maudheim expedition, reacted to climatic change (Liestøl 1953).

Vatnajökull is not, however, the largest glacier that has begun to react to climatic change, for also the glaciers of Spitzbergen (Ahlmann 1933, 1935, Glen 1939) and Greenland (Bauer 1955) are currently shrinking. In studying the present shrinkage of glaciers, Thorarinsson has gone through, *inter alia*, many reports of Greenland and concludes that »recession and thinning have for some decades characterized the large majority of the Greenland outlet glaciers and the margin of the inland ice» (1940 p. 147). From Thorarinsson's study it becomes quite clear that the shrinkage of glaciers is a universal phenomenon, which, with the possible exception of Antarctica, is evidenced by the present behavior of all the glaciers on earth.

Size is not the only basis for comparing the behavior of glaciers and studying the results. A different and perhaps better basis for comparisons is a grouping of glaciers into different types. Many kinds of classifications have been made; some have been pretty general, while others have been developed to a high degree (e. g. Hobbs 1911, Priestley and Wright 1922, Ahlmann 1933, Drygalski-Machatschek 1938, v. Klebelsberg 1948). Most of them are based on the morphological characteristics of glaciers and especially the placing of some among intermediate types has perforce been subjective. In order to reach firmer ground in the morphological grouping of glaciers, Ahlmann has revised his morphological type classification (1940) by carrying out analyses to determine how the area of different-type glaciers is distributed percentably at different height intervals. The results of the analyses have been presented as area distribution curves (op. cit. p. 198). When the curves thus obtained are compared, it becomes clear that ice caps and ice sheets have the same kind of curves as Hoffelsjökull and the other subsidiary glaciers of Vatnajökull. In other words, they belong to the same glacier type as Vatnajökull as a whole as well as other ice caps and Greenland. The curve shows their area as expanding toward the broad plateau of the summit. The curve of Ahlmann's transection glaciers, including e. g. the Lilliehöök and Löwenskiöld glacier in Spitzbergen, the Yakutat glacier in Alaska as well as other »*Eisstromnetze*», starts higher, rises convexly to its maximum already at the 6—8 interval and falls thereafter steeply below the level of its starting point. The curves for valley glaciers are similar, with the culmination points being situated at different intervals. The curve of the great Aletsch glacier resembles the ice cap curve most, whereas the Piedmont-glacier curve differs from it most radically, as it culminates at the lowest interval. The ice mass of Piedmont glaciers is thus situated quite differently from ice sheets and ice caps. Its expanded part is so large that it acts independently and melts at the same time rapidly. Since the main part of the ice mass of a Piedmont glacier is situated in the ablation zone, a rise in temperature does not alter the relation between the accumulation and ablation areas of such a glacier as significantly as in the case of ice sheets and ice caps — unless the melting reaches a stage where the Piedmont glacier loses contact with the ice streams feeding it. A change in temperature affects the behavior of a glacier the more forcefully the greater the relative area of the higher parts of the glacier (Ahlmann op. cit. p. 203). Accordingly, the recent amelioration in climate must be reflected most plainly in the activity of ice sheets and ice caps, for their balance has been disturbed most.

Rapid melting ought, therefore, to characterize the marginal zones of all glaciers today whose accumulation areas have recently shrunk, i. e. ice sheets, ice caps and transection glaciers. It is, of course, not evident identically in all glaciers, inasmuch as both climatic and topographic circumstances affect the behavior of each ice mass. This becomes plain upon studying the literature on the glaciers of our times.

The dead volcano Beerenberg, on the island of Jan Mayen N-NE from Iceland, provides the base for an ice cap some 70 km^2 in area but nevertheless thin (Odell 1939, Jennings 1939, 1948). From the climatic snowline, which in 1938 was situated 914 m above sea level on the southern slope and 636 m on the northern slope (Jennings 1948 p. 180), several thin but broad outlets extent seaward. Some of them extend as far as the sea, but others terminate on land. In front of the latter are two arcuate end moraines, from which the glacier has already retreated. When Odell visited the island in 1933, he noted dumps in the farther end moraine, and in 1938 Jennings found an ice core in the inner end moraine at numerous points. Ice was found also buried under lateral moraines, and in front of a number of ice tongues there occurred detached parts of stagnant ice. Furthermore, the termini of the tongues were in many places covered with drift (e. g. the eastern Weyprecht glacier). From Jennings' description it is evident that the glacier is rapidly shrinking, that dead ice occurs along its borders and that the marginal parts of the tongues terminating on land are beginning to be covered with moraine. The glacial drift undergoes the same kind of treatment, therefore, as in Iceland.

The glaciers of Spitzbergen belong to the subpolar type. They receive most rain in West Spitzbergen, where the climatic snowline is situated 200—800 m above sea level, since moist winds blow over the islands from the Atlantic. In the northeastern part of the island there is less precipitation and the climatic snowline is there situated 350—550 m above sea level (v. Klebelsberg 1949 pp. 570—571). Inasmuch as the climatic conditions become more rigorous toward the northeast, the glaciers of North East Land are more uniform and larger than those of West Spitzbergen (Ahlmann 1933).

In West Spitzbergen there have evolved transection glaciers and small ice caps, and the highest and also coldest, northeastern part of the island is covered by a uniform ice dome. From their firn areas there jut out seaward outlet glaciers, which in places calve into the sea and in other places terminate on land. The ice flow in the outlets is rapid and causes changes also in the position of the margin. The borders of such ice tongues have become known by virtue of numerous studies (e. g. De Geer 1910, Lamplugh 1911, Hoel 1916, Tyrrell 1927 and Gripp 1929.) According to Gripp's observations the movement of the ice takes place in the marginal zones of the tongues along parallel shear planes. These planes lift glacial drift upward, for the base of the ice mass turns uphill as soon as the margin encounters the slightest obstruction. The advancing margin itself helps form obstructions by building up in front of it push moraine out of deposits in the frontal terrain. In addition, basal load is squeezed upward along crevasses in the glacier. Obstructed flow thus plays a notable part in the flow mechanism of the outlet glaciers of West Spitzbergen. In consequence of it there evolves *kuppige Grundmoränenlandschaft* (termed ablation moraine in the foregoing) in extensive zones. Since the action of meltwater is slighter than in Iceland, topography like this generally survives. The meltwaters largely flow through englacial tunnels and out of their sorted material sandurs are produced beyond the glacier margin. Their occurrence in West Spitzbergen, where volcanogenic glacier outbursts do not figure, provides evidence that sandurs evolve as a result of normal glacier drainage.

Gripp's precise observations make it manifest that the glacial drift deposits of West Spitzbergen correspond in a high degree to those appearing along the southern margin of Vatnajökull. This similarity has attracted the attention e. g. of Todtmann (1932) and Hoppe (1951). The differences are mainly such that there is a greater abundance of eolian sediments on the glaciers of Iceland, the washing of drift is stronger there and basal load is released from the ice there sooner than in West Spitzbergen.

The glaciers of North East Land are uniform ice caps, separated by iceless zones. The caps lack nunataks and their surfaces form domes reaching up to the 700 m level. Their margins either terminate on land or calve into the ocean. On land they form thin sheets of ice covered with glacial drift, many of which have become detached from the glacier proper and melt in place (Glen 1937). The disappearance of the glaciers is promoted by the undermining effect of meltwater. Thus Groft relates having found, in the winter of 1935, a crevasse in an ice cap in North East Land down into which a man could descend 70 feet. »At the bottom was an elaborate and strikingly beautiful series of caverns, with a considerable lake of water. Here was the bottom melt in full operation — a state of affairs which was very different from that on the surface» (1937 p. 249). Fracturing of the thin marginal parts of the glaciers is so common that Glen (1939) foresees the ice caps of North East Land breaking up into numerous separate glaciers in the future. According to Ahlmann they are inactive, »starving» glaciers, whose accumulation, ablation and flow are all slight. In this respect they differ from the active, maritime glaciers of Iceland. In times past the glaciers of North East Land used to be more active and built up end moraines (v. Klebelsberg 1949 p. 570).

The conditions prevailing at the southern margin of the Barnes ice cap, which the expedition of the Arctic Institute of North America investigated in the summer of 1950, afford an illuminating picture of the way glacial drift rises up in the ice margin. Shaped like an irregular ellipse oriented NW-SE, the Barnes ice cap covers an area of about 5 900 km^2 in the eastern central part of Baffin Island. Its cupola-like center rises to an elevation 1 120 m above sea level, while its margins lie at the 450 m level. According to meteorological and glaciological observations (Orvig 1951, Baird 1952) the Barnes Ice Cap is subpolar or high polar and receives very little precipitation. After investigating the southern margin of the glacier, Goldthwait (1951) decided that the margin had at certain points remained stationary perhaps for centuries but had receded in our own times at other points. The slope of a retreating ice margin was slighter than that of a stationary one. So much basal till rose up to the surface of the ice margin in the latter case along the numerous thrust planes that the ice was buried under surface drift. This occurred to a height of 100—200 feet, measured from the margin. In spots where the drift layer was three feet thick, it retarded the melting of the ice and formed ice-cored moraine, separated from the main body of the glacier by the more rapidly melting trough. According to gravimetric measurements (Goldthwait op. cit. Fig. 3) the toe of the glacier rests on a base rising outward along a gentle slope and the thickness of the ice at the point where dirt begins to appear on the surface is about 150 feet or 50 m. Goldthwait's investigations reveal that obstructed flow takes place also at the edges of cold ice caps. Since the surface of such glaciers undergoes less washing than that of the glaciers of Iceland, the drift cover of the border crest grows thicker and the whole process is more plainly evident than in Iceland. It is interesting to note that according to Goldthwait's conception the rise upward of drift on a slant plane occurs also when the margin of the glacier terminates as a steep cliff in water (Profile B, Fig. 3), i. e. precisely as at Hagavatn in Iceland. Ward too (1952) had made similar observations regarding both transversal moraines and the margin of glaciers terminating in a lake (Fig. 13 and photo Fig. 4). He has further noted that ice buried under drift outside the visible ice margin is likely to remain unmelted for long periods of time, because of the cold climate (p. 21). The drainage of the Barnes ice cap differs from that of the Icelandic glaciers, for on the basis of Ward's observations the meltwater forming in the summer flows across the surface of an unbroken glacier up to the margin rather than subsurficially. This is because, Ward propounds, the temperature of the ice is usually below the pressure melting point, causing water penetrating into the ice to freeze (p. 11). Likewise, crevasses opening in the ice also become sealed. On the surface of the ice behind the border crest shallow lakes form and in places in front of the glacier dammed-up marginal lakes.

Greenland's great ice sheet (Hobbs 1911, v. Klebelsberg 1949, Woldstedt 1954) naturally affords many points of comparison, for at its various margins there occur, as a consequence of both topographic and climatic conditions, various kinds of activity. The vast outlets calving into the sea rank among the most rapidly moving glaciers on earth, whereas the ice margin of northern Greenland, which in places terminates on a high, flat plateau and elsewhere forms a transection glacier type of margin in the mountains, represents a high polar, inactive glacier. As a whole, the continental ice sheet of Greenland, according to the latest calculations (Bauer 1955), is 1.726×10^6 km^2 in area and 2 135 m in average height, with the elevation of the climatic snowline averaging 1 390 m above sea level. The snowline divides the glacier in such a way that the accumulation area comprises 83.5 % and the ablation area 16.5 % of the total. According to Bauer's calculations, the budget of the glacier is negative, for the accumulation (446 km^3) does not make up for the loss caused by ablation (315 km^3) and calving (215 km^3). The annual loss amounts on the average to 100 km^3 of water.

Even though the ice sheet of Greenland as a whole is wasting away, its great accumulation area does not yet reflect it. Its even field is without crevasses, moraines or nunataks (e. g. Nansen 1892, Koch und Wegener 1911, 1930). The effect of the margin is felt farthest toward the center at the points where rapidly moving ice streams extend to the sea. In the western part of Greenland the crevassing caused by them occurs as far as 145 km from the ice margin (Quervain 1920). Possibly the terracing to be observed in the marginal parts of the ice sheet (Woldstedt 1954 p. 38) is caused by the fact that the relief of the base has begun to affect the form of the glacier. One special consequence of this irregularity developing is the formation of large lakes on the surface of the ice. Only in the marginal zone does glacial drift begin to appear on the glacier. It forms lateral and medial moraines, which are flushed by water in transit, eventually evolving into block fields on the ice (Chamberlin 1897, p. 239). On the glacier margin the amount of drift increases, for basal drift rises up shear planes and crevasses to the surface. Its occurrence has been described by many investigators (e. g. Chamberlin 1894a, 1895, 1896 and 1897, v. Drygalski 1895, Koch and Wegener 1911, 1930). According to Chamberlin's observations (1895, pp. 572, 675) load rises up along shear planes inclined forward, especially at points where the motion of the glacier takes place uphill. An obstruction is likely to be formed by push moraine accumulated by the glacier itself (p. 675). Also more slowly moving sheets of ice form an obstruction to those farther behind, forcing them to thrust forward obliquely (p. 677). Movement up shear planes takes place by jerks. According to the observations of Koch (1911 p. 438) there has evolved as a result of such movement the broad ablation area of Germanialand in NE-Greenland, where numerous clay towers and pyramids had risen on the ice surface.

Also Battle (1951) emphasizes the importance of thrust planes to ice movement in the Fröya glacier of NE-Greenland. This glacier, Ahlmann (1941) points out, is one of the best representatives of the high polar or inactive glacier type, for its regimen is characterized by very slight accumulation and equally slight ablation, combined with a low temperature. Similar glacierization conditions prevail also at the W-margin of the Greenland ice sheet at Eqe, where the glacier terminates on a relatively even high plateau. As Boyé and Cailleux recently (1954) noted, the ice margin there is surrounded by a broad ablation zone, in which both radial and transversal crevasses occur. One transversal crevasse was situated 200 m upstream from the outer border of the ablation belt and constituted the true border of the ice sheet, separating as it did the ablation moraine-covered part of the glacier from the glacier proper. Into the channel formed by this crevasse, drift avalanched from the surface of the melting ice mass. At the bottom of the crevasse could be seen morainic boulders. When Boyé and Cailleux measured the ice crystals, they were unable to ascertain from what depth the crystals had come »because of the overthrusting and folding of the ice». This interesting description reveals that the margin of the Greenland ice sheet at Eque had become buried under surface drift raised by obstructed flow and cut off from the glacier.

The development at Eqe had thus led to precisely the result represented by Rich (1943). He had reached the conclusion, in dealing with the matter theoretically, that the conditions prevailing at the ice sheets favor the burial of the ice under glacial drift. The thin and drift-covered outer edge of a glacier forms an obstruction to advancing ice, forcing it to slide upwards at a slant, which leads to rising up of drift along thrust planes toward the surface. As the ice margin thins from melting, drift is brought into view and it produces a border crest, separate from the main body of the glacier by a trough. Rich's drawing (op. cit. Fig. 2) resembles so greatly those presented by Goldthwait (1951 Fig. 2) and Ward (1952 Fig. 7) that all three might illustrate the same ice margin at different stages. The turning upward of the ice margin, which Hobbs considered characteristic of mountain glaciers, thus proved to be also characteristic of ice caps and the ice sheet of Greenland. Previously the same phenomenon had been encountered in the Piedmont glaciers of Alaska (Russell 1895, Tarr 1909, Tarr and Martin 1914, Washburn 1935), where the ice margin is covered by drift in a broad zone. Adding to this the fact that in cirque glaciers (Clark and Lewis 1951) the movement of the ice takes place up slant shear planes, it may be affirmed that obstructed flow occurs in the present margins of different glacier types. According to the observations of Ward (1952, p. 12), the drift raised by this motion is released from the ice in two ways: in response to ablation and in the case of active ice gliding over less active or dead ice. Both processes occur when drift is released from an active shear plane, but only the former upon the melting of a stagnant shear plane. To what extent upraised drift remains on the surface depends largely on the strength of the surface washing. In Arctic glaciers, where surface washing is less abundant than in more temperate conditions, surface drift has greater possibilities of forming such thick layers that melting of the underlying ice is effectively prevented (Sharp 1951). On vigorously washed glaciers, including those of Iceland, border crests do not develop as fully as in more rigorous climatic conditions. The structure of the margin nevertheless distinctly shows the same process to prevail.

From the southern hemisphere Nicols and Miller (1951) have depicted the same kind of treatment of glacial drift at the Ameghino glacier as has herein been presented in connection with the glaciers of Iceland. The Ameghino is an outlet approximately 15 km long and 2 km wide of the transection glacier situated on the western side of Lago Argentino in the Andes (50°25′—51°25′ southern latitude). On the surface of the glacier there is to be observed e. g. thrust planes generally inclined toward the center, transversal crevasses and places where the glacier has advanced on top of stagnant blocks of ice. In front of the glacier there has formed a lake dammed by frontal moraine; and the terminus of the glacier is submerged in the lake. Moreover, there is an abundance of roundstone till in the glacial drift; and this is regarded by the researchers as one proof that the glacier has advanced over the outwash plain formed at the margin and from it taken material to be deposited in the till. Thus in the case of the Ameghino glacier formations analogous to those in Iceland are met with.

The glaciers along the southern coast of Alaska, already referred to, have become known as representatives of a certain Piedmont type. Their multi-branched accumulation areas collect ice into extensive low-land glaciers, which have evolved below the snowline (Sharp 1951). The largest in size is the Malaspina glacier, whose ice bulk comprises an area of 3 840 km^2 (v. Klebelsberg 1949 p. 494). On the basis of seismic soundings the thickness of the ice in the middle part of the glacier is 1 130—2 075 feet (376—690 m) and its relatively even floor is situated below sea level, rising gently toward both the mountains and the sea (Sharp 1952). In the middle part of the glacier the ice crosses a gently sloping sill in moving seaward. Behind this sill Sharp sank a tube 1 000 feet (330 m) long in order to determine the flow mechanism of the ice. The results up to now (Sharp 1953 a and b) indicate that »the surface ice appears to be carried along by the flowing ice beneath in a manner contrary to the mechanism of extrusion flow» (Sharp 1953 a p. 184). And in another report Sharp (1953 b) writes: »The initial results are certainly not consistent with the mechanism of extrusion flow as expounded by Demorest». At the glacier margin the movement of the ice turns upward and so much drift is brought to the surface up the shear planes that the entire border zone is buried under it. The drift accumulated at the margin receives addition material from medial moraines, which spread out on nearing the edge tens of kilometers (Tarr 1909, Tarr and Martin 1914). As the Malaspina glacier is located in a region of heavy precipitation and as, furthermore, it lies wholly in the ablation area, its drainage is abundant. The water is discharged under pressure from the edge of the glacier, and it carries with it much glacial drift, from which an extensive outwash plain has been built between the glacier and the sea. In flowing across this plain the streams branch out frequently and continually change course. They behave, in other words, in the same way as the glacier-fed rivers of Iceland, with the exception that they go through only a single phase, whereas in Iceland the water drained from glaciers undergoes at a number of places three phases. The action of meltwater from the Malaspina glacier thus is about the same as during the formation of the Icelandic sandurs. The flow of water, which according to the observations of Russell (1893) and Tarr (1909), occurs both englacially and subglacially, is augmented from ice-dammed lakes located on the northern side of Malaspina between the glacier and the bordering mountains. Here too we see close similarities between the glaciers of Iceland and southern Alaska.

The drainage of the great ice sheet of Greenland naturally varies at different parts of the glacier. Meltwater is abundantly discharged from the ice tongues terminating in the fjords of the southern part of Greenland, and from the glacifluvial material carried by it outwash plains are built up at the heads of the fjords. The meltwater also washes the moraine released from the ice and re-sorts it to the extent that original moraine deposits are in many places a rarity. At the margins terminating in the interior much less meltwater is met with. It vanishes from the surface of the ice into crevasses and issues forth from the glacier tunnels at the edge. According to Nansen's observations (1892) a heavy flow of water continues out of the tunnels even after the air temperature has been below the freezing point for several months. The glacier thus seems to be melting subglacially even in the wintertime. Woldstedt (1954 p. 40) considers the reason to be either heat stored up in the land or friction caused by the movement of the ice. A similar phenomenon is encountered on North East Land, as related on p. 116. Battle (1951) tells of running across a tunnel in the Skille valley glacier in northeast Greenland at the bottom of which flowed a stream. When the tunnel was explored to a length of 500 m, it was ascertained that the water came into the tunnel from a crevasse overhead, which had opened up in the ice under a medial moraine. The meltwater forming on the medial moraine on top of the glacier thus first collected in the crevasse and then flowed subglacially through the tunnel to the edge of the glacier (comp. p. 37). At the point where the crevasse was situated there was 76 m of ice above the tunnel. Similar empty tunnels had previously been met with in the glacier of Malaspina (Russel 1893) and at Skeidarárjökull after a glacier outburst (Nielsen 1937). After the water has ceased to flow, the tunnels, however, soon close up.

In regard to abundance and activity of meltwater, the glaciers of Iceland compare most nearly with the glaciers of southern Alaska and the southern part of the Greenland ice sheet, where the meltwater flows subsurficially and is discharged from the mouths of glacier tunnels. The drainage of the

Icelandic glaciers differs most from that of subpolar and high polar ice caps, which by and large takes place superglacially. The ice margins terminating in shallow water of marginal lakes, such as those described from Hagavatn (p. 90), the edge of Hoffelsjökull (p. 90) and Hvitárvatn (Nørvang 1937) in Iceland, form the same kind of submerged outlets also in Alaska (Russell 1893, Hobbs 1911 p. 180) and Norway (de Boer 1949).

A common feature of the glaciers considered in the foregoing is that they have formed on land and their margins terminate either on land or in shallow marginal lakes. In addition, they are all currently melting away. In spite of these characteristics in common, however, distinct differences appear both in their behavior and the treatment undergone by drift. They become most clearly manifest when high polar ice caps are compared with temperate glaciers. The former are inactive and slow of movement, and wastage takes place more by evaporation than melting. The latter are active and rapid of flow and melt simultaneously both on the surface and at the base. In the case of both types glacial drift rises up to the surface as a result of obstructed flow, but these accumulations are better preserved in the marginal zones of high polar glaciers, forming on melting pitted end and ablation moraine. On the surface temperate glaciers the drift wanders more easily. It is flushed down into depressions, crevasses and in front of the glacier edge. Since, furthermore, glacifluvial activity is stronger in temperate glaciers, the glacial drift is altered into glacifluvial matter, which is deposited in front of the glacier margin to produce extensive outwash plains. Such deltas sometimes evolve also on top of the thinned-down margin of a glacier.

SUMMARY

1. The glacierization of Iceland is a result of the combined effect of two components: topography and climate. The largest glaciers on the island are ice caps, whose firn areas are surrounded by ablation zones. In the interior broad, short lobes occur on the glacier margin; but along the southern coast long outlet glaciers extend nearly down to sea level. The behavior of the glacier margin depends on the pressure prevailing at any given time on the main body of the glacier. In recent decades the glaciers have been melting fast. This can be noted in the ablation area in the thinning of the ice and in a cessation of the glacier flow. Obstructed flow is the most obvious kind of movement in the present margins of the Icelandic ice caps.

2. The moving ice carries drift in abundance to the marginal zone. This load is divided into wind-blown detritus, volcanic products, surface moraines, basal load, washed material, and crevasse fillings. During

transportation the position of the drift varies. The drift deposited on the surface of the accumulation area is rapidly buried under snow and comes into view only in the ablation area. Here it produces medial and lateral moraines as well as upturned ash layers and dirt bands. The medial and lateral moraines contain much disintegration material avalanched onto the ice surface, but some medial moraines originate from the ice-covered peaks of the glacier floor and contain only erosion products. The superglacial drift retards the melting of the underlying ice and adds to the irregularity of the glacier surface. At the same time the various components become mixed up. Thus e. g. the glacial drift receives pollen and spores from the recent eolian detritus.

Obstructed flow also forces basal load up along thrust planes to the ice surface in the marginal zone. The uplifting process takes place in various ice border types. Load rises up in the ice at least 30—50 m. On coming into sight, it is wet, filling the transversal crevasses between upthrust sheets of ice. The crevasses dip about 30°—40° towards the glacier. Thick accumulations of load remain elevated and form transversal moraines, as the ice surface thins down faster between them. At last, the border crests and other superglacial accumulations are detached, along with their ice cores, from the active glacier. If the surface washing is vigorous, border crests of long duration cannot form, for the material is flushed from the heaps into hollows and crevasses.

The hollows are liable thereupon to collect so much drift that they in turn melt more slowly and end up as mounds.

At the same time as the glacier thins down on its surface, melting takes place also at bottom, for basal load cuts loose from the ice already subglacially, forming basal till. In general the till detached from the glacier remains undisturbed in place, preserving the fabric given it during transport, but at some points, however, the ice squeezes up till into the crevasses and in front of the margin.

3. The moraines released from the ice have been divided into ground, radial, ablation and end moraines. Ground moraine covers wide areas in the gently undulating highland of the interior, where it constitutes an accumulation evening out irregularities of the bed, though still reflecting them. The occurrences of ground moraine along the southern margin of Vatnajökull are small in area. Since the nunataks and bordering mountains exposed out of the ice are usually bare, it is conceivable that the basal load does not rise high enough in the ice to form thick accumulations on the mountains. The best conditions for the creation of ground moraine exist when the ice melts in a stagnant state and does not contain much surface drift. If, furthermore, the melting takes place by evaporation, the ground moraine preserves also its own relief. At a glacier margin that oscillates

and forms meltwater in abundance, the accumulations of basal till are deformed.

Radial moraines form out of basal till in the marginal crevasses. The till is squeezed up from underneath the ice. A necessary condition for the pressing up of till is that it must have become detached from the ice already at a subglacial stage. Small radial moraines originate also when blocks of ice fall off the glacier margin into mushy drift, pressing it into the spaces between them. Radial moraines occur in limited areas only.

Ablation moraines evolve on the margins of the glaciers of Iceland nowadays quite generally. They are met with along the sides of outlets, in front of medial moraines and everywhere that obstructed flow lifts sufficient basal load up to the surface of the glacier. In the trough between the glacier margin and arcuate frontal moraine there almost invariably occurs ablation moraine. The formation of ablation moraine takes place everywhere that the ice melts underneath glacial drift. The ice first wastes away in place where there is little surface drift. Into depressions thus formed, melted rapidly by water, there slides such an abundance of drift from the surface that the original holes collect a thicker accumulation than remains on the last lingering ice. As the drift at the same time becomes sorted and re-treated, it loses its original structure. It contains deformed sediments mixed up with till.

Ablation moraine forms abundantly when the glacier melts rapidly and thins down and when the ice contains sufficient glacial drift. As a consequence of thinning, the marginal parts of the glacier turn stagnant and detache themselves from the active glacier. Stagnation of ice may also occur in places where the ice is free of drift. That is why the occurrence or absence of ablation moraine cannot be used as proof in explaining whether the glacier had been active or turned stagnant in its last phase. The forming of ablation moraine is generally concentrated along the glacier margin. If the marginal zone retreats continuously, the areas of ablation moraine are likely to spread considerably, without, however, signifying that any large parts of the glacier had simultaneously turned stagnant.

End moraines have developed most prominently in front of the southern outlets of Vatnajökull, but they are to be met with elsewhere as well. Their distal sides border on great sandurs and their proximal sides either on marginal lakes or ablation areas. In only a few places does the glacier still extend to their proximal slopes. The great end moraines have formed during the earlier stage of development, when the ice margin was thicker and extended farther than now. It formed push moraines out of the frontal deposits, thus creating for itself conditions whereunder the ice border had to advance uphill. The consequence of this was an increase of obstructed flow. The push moraines thereupon became covered by both dump moraine and basal load. The rise of basal load occurred to a much greater extent than now at the time when the ice margin was in active motion at the site of the end moraines. As this activity has nowadays changed to passive melting, end moraines form only in exceptional cases. They are modest in size and accumulate at a glacier margin when it advances or remains stationary. As a consequence of the different ways in which end moraines form, their material and structure vary considerably. In them basal till, folded sediments from the frontal area, and surface drift occur as alternating and also intermixed strata. The knob and kettle topography of the great arcuate end moraines verifies that in the end moraines there had been buried ice in abundance.

4. The wasting away of the ice is promoted by abundant rainfall, especially along the southern seaboard of Iceland. As, moreover, much meltwater forms, great quantities of water are discharged by the glaciers each year. The flow of water takes place almost wholly subsurficially. At least part of the water streams through tunnels with a natural slope. During glacier outbursts the flow takes place subglacially under hydrostatic pressure, but at other times the pressure is slight. The water collects into marginal lakes, which in places extend up to the crevasses and in other places continue under the ice or inundate the thin ice margin. The water of marginal lakes melts the ice most along the waterline, with the result that at the bottom there remains ice buried under sediments. From the lakes the water flows into swift rivers which branch out vigorously and constantly change their course. The activity of the meltwater thus commonly has three stages.

5. The glacial drift which comes into contact with the streaming water changes in character, being termed glacifluvial material. Part of this treatment takes place on the surface of the ice, part subsurficially and part outside the glacier.

The deposits forming on the ice are sometimes stratified, sometimes accumulations of fine and coarse sand with scattered stones and more weakly sorted parts. Crevasses, in particular, and the fosse behind the border crest serve to collect these sediments. During the final stage of melting these sediments usually became mixed with other surface drift and occur as lenses in ablation and end moraines. Deposition of glacifluvial material likewise takes place on thin submerged sheets of ice. These deposits deform into pitted terrain when the ice melts from underneath.

Glacial drift treated by subsurficial streams constitutes the main part of the glacifluvial deposits. During glacier outbursts the water issues from the tunnels with such force that it carries all the drift outside the glacier. In such circumstances the material forms an esker in front of the mouth of a tunnel. Ordinarily the glacifluvial drift spreads out into outwash plains. The coarsest part accumulates nearest the glacier, while the finest is carried farthest. Stones and boulders in such a deposit are orientated so that their longitudinal axes lie perpendicular to the flow. Flat-topped boulders are turned so that their upper surfaces lie at a rising angle downstream. As the central cones of the outwash plains generally lie on a higher level than the present glacier tunnels, the glacifluvial activity was stronger during the formation of the sandurs and more glacifluvial drift was discharged by the glaciers than at present.

The water drained from the glaciers nowadays flows out gently and without pressure. It carries only fine-grained material, which deposits in the marginal lakes. The coarser glacifluvial material accumulates subsurficially, forming *inter alia* tunnel eskers. Evidently, the deposition of glacifluvial drift takes place also in marginal lakes extending under the ice, which represent the first sedimentation basin for the material of subsurface streams. The sediments resemble the washing on the glacier surface. Sometimes they are stratified and often they are mixed with drift sliding down off the ice. If the sedimentation takes place off the ice onto a free base, the sediments level it out; but if they accumulate on top of ice at the lake bottom, their layers are later deformed.

The strongest treatment currently given glacial drift takes place outside the glacier border. The braided rivers emerging from the marginal lakes have so strong current on account of the steep gradient that they rapidly erode the drift within their sphere. The moraines formed previously are worn off and their material changes under the treatment of the streams. The earlier delta planes likewise erode, with new beds and delta isles in between being created. The erosion is strongest in the proximal parts of the sandurs, from which the marginal lakes prevent the acquisition of additional load. The delta plains subject to two-sided erosion are transformed into esker-line erosion remnants. Their primary stratigraphy represents typical sandur structure. The sides of an erosion esker are at first vertical, but in time they begin to slope less steeply as the material dries and collapses. If such erosion eskers are preserved, they form an esker type of their own.

6. The accumulation of glacial drift described in the foregoing belong to two genetic stages. The one corresponds to the present behavior of the glaciers and the other represents prior conditions. In addition, the drift also contains fragmentarily material reflecting still earlier stages of glacierization. The conception of the course of development in southern Iceland is presented in the accompanying table (p. 127).

7. The treatment undergone by glacial drift on the glaciers of Iceland corresponds to that on other temperate glaciers. It is characterized by active behavior of the ice, heavy melting simultaneously on the surface and at the base of the glacier, abundance of meltwater and its subsurface

Formations	Glaciers	Meltwater	Sea	Climate	Date
Ablation moraine, ground moraine, sediments of marginal lakes, erosion and tunnel eskers. Radial and end moraines rare	Melting. Margin often stagnated Balance negative	Abundant, but occurring in three stages on account of marginal lakes' forming. Volume of water in icedammed lakes diminishes	At 0-level	Warm and rainy	Present stage
Great end moraines evolve. Troughs hollowed out under ice margin	Active. Margin extends to end moraine. Balance ±0	Abundant, subject to pressure. Discharged in one stage. Much water in ice-dammed lakes. Few marginal lakes	At 0-level	Rigorous	Latter half of 19th century
Betula peat on Breidamerkurjökull. *Betula* stems on Hoffelsjökull. *Betula* peat at Hoffel and Laugardalur	Much smaller than at present	?	Perhaps +4 m (Bárdarson 1910, Thorarinsson 1951). Does not extend to Hoffel nor Laugardalur.	Warm and humid	Postglacial climatic optimum
Ericaceae-Salix pollen at Hoffel and Laugardalur	Larger than at present. Margin perhaps at great end moraines	?	Near the swamp of Laugardalur (+15 m). Does not extend to Hoffel (+30 m?)	Colder than now perhaps dry	Preboreal
			In northern Iceland +40—50 m (Thorarinsson 1951) High water stage?		Gotiglacial
Moraine at Hoffel and Laugardalur. Striations on coastal plain	Larger than at present. Margin extends at least to present shoreline	?	Low	Cold	Last general glacierization in South Iceland
Sterile mollusc clay on Hoffelsjökull, Breidamerkurjökull, Skeidarárjökull and Öræfajökull. The sterile clay of Kvisker and the mollusc clays of Gullfoss and Ellidá originate? Shore terrace evolves	At minimum size	Strong glacifluvial activity in Icelandic interior. »Palagonite formation» produced?	As far as coastal terrace (+100 m). Cold and transgressive	Warm	Interglacial or interstadial stage
Moraine and striations, e. g. under interglacial strata. Submarine valleys evolve?	At maximum size	Glacifluvial activity on submerged coastal plain	Low, at level of submerged coastal plain (—200 m?)	Cold	Ice Age

flow. Previously produced end moraines, from which the ice margin has receded, constitute natural obstructions to the flow of meltwaters and cause it to undergo three separate phases. Local conditions also bring about an augmentation of the quantity of drift. As a consequence of its abundance, ablation moraines are generally met with. All in all, the glacial drift of Iceland affords a coherent picture of the kind of formations nowadays evolving in the marginal zones of rapidly melting and simultaneously active, temperate ice caps.

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FIGURES No:s 1—32

IN

PLATES I—XVI

PLATE I

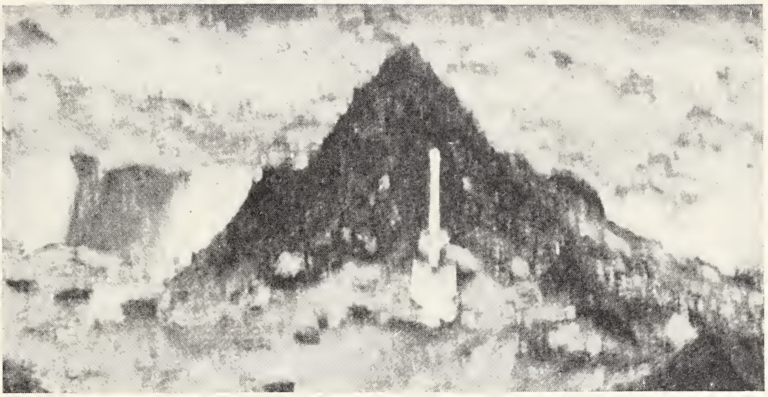

Fig. 1. Ice-cored dirt cone on Hoffelsjökull. The length of the spade is 75 cm.

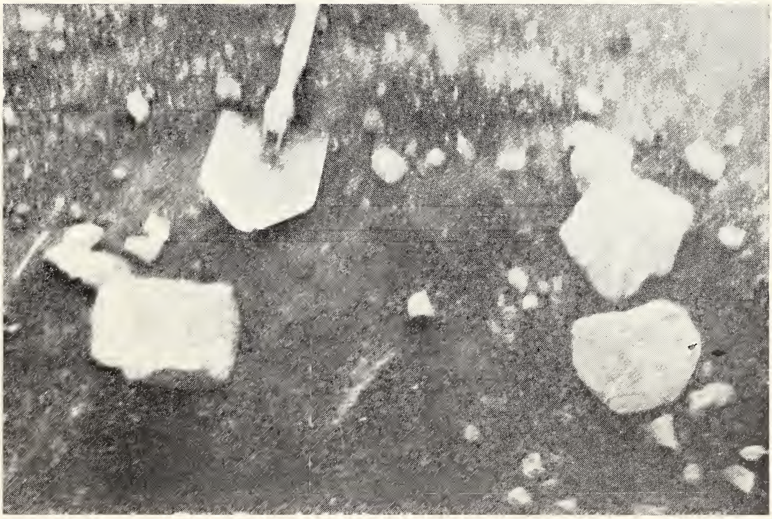

Fig. 2. Glacial drift in a medial moraine on Hoffelsjökull. The rock fragments are angular and lack striations.

PLATE II

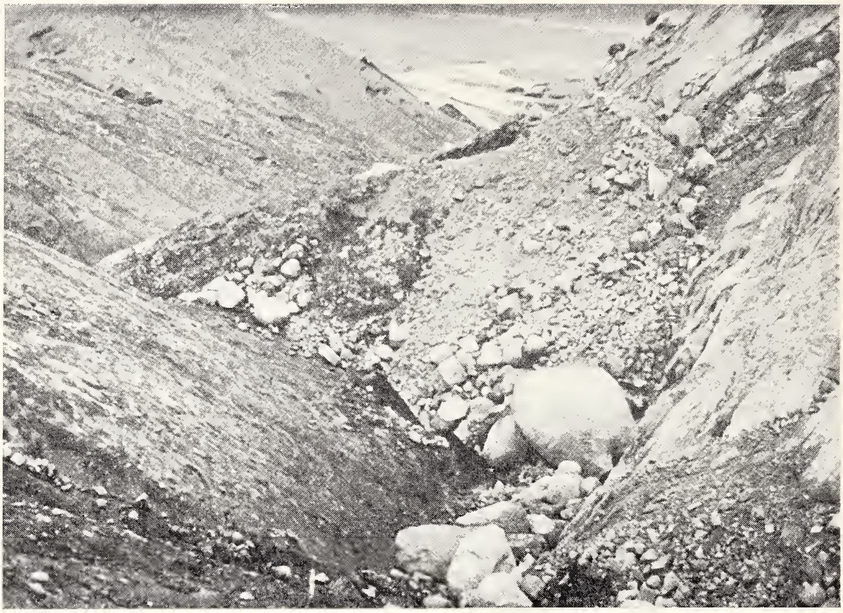

Fig. 3. Lateral moraine along the western side of Skaftafellsjökull (left) is accumulated on the ice. It consists mainly of material from the rock Beneath slope (right). the moraine an ice core.

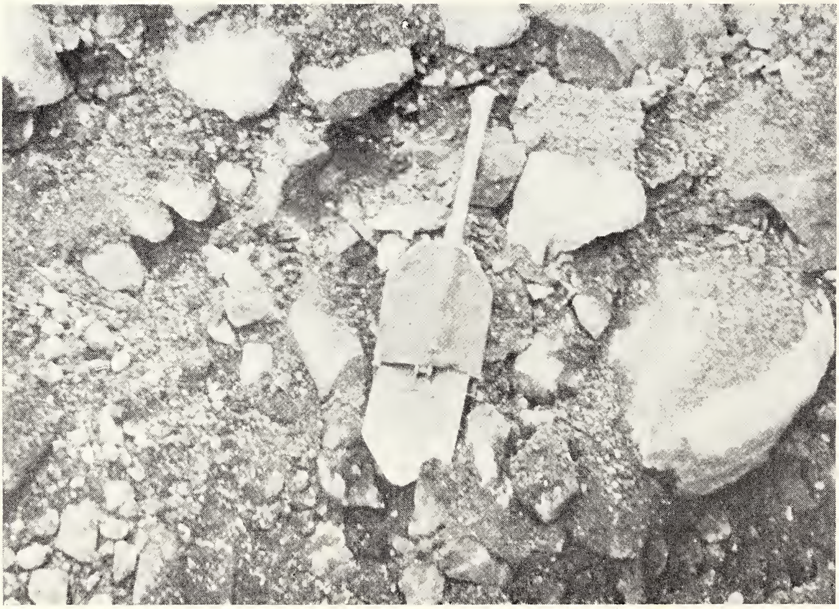

Fig. 4. The lateral moraine seen in the former picture grows thicker downstream. Near the terminus of the glacier it forms a mantle two meters thick on the stagnant ice.

PLATE III

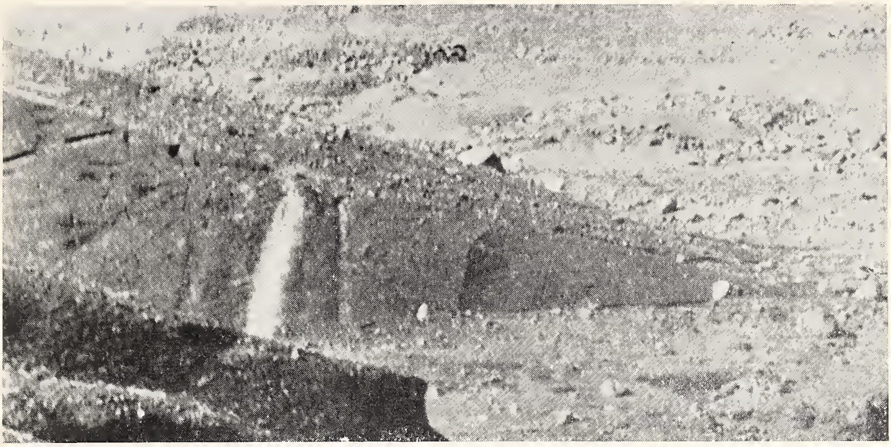

Fig. 5. The snout of Breidamerkurjökull (left) forms a thin wedge upon the ground moraine. Dark stretched ovals in the icewall (top left) are hollows in the ice and they are lengthened parallel to the ice movement.

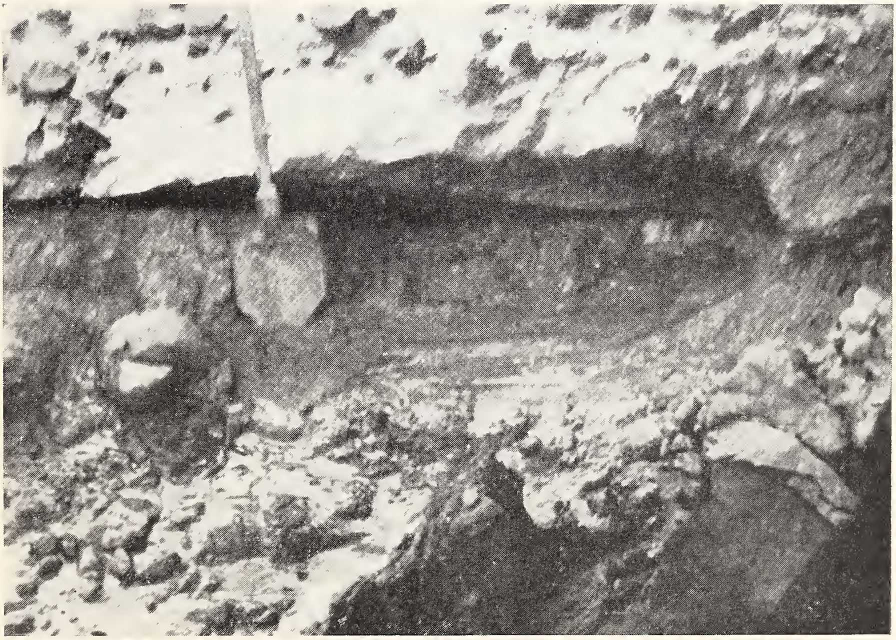

Fig. 6. A transversal crevasse 30 cm in width in the snout of Hoffelsjökull is full of water-soaked till. The tip of the spade rests on the underlying ice while the overlying ice sheet begins behind the lower part of the spade handle. The length of the spade is 75 cm.

PLATE IV

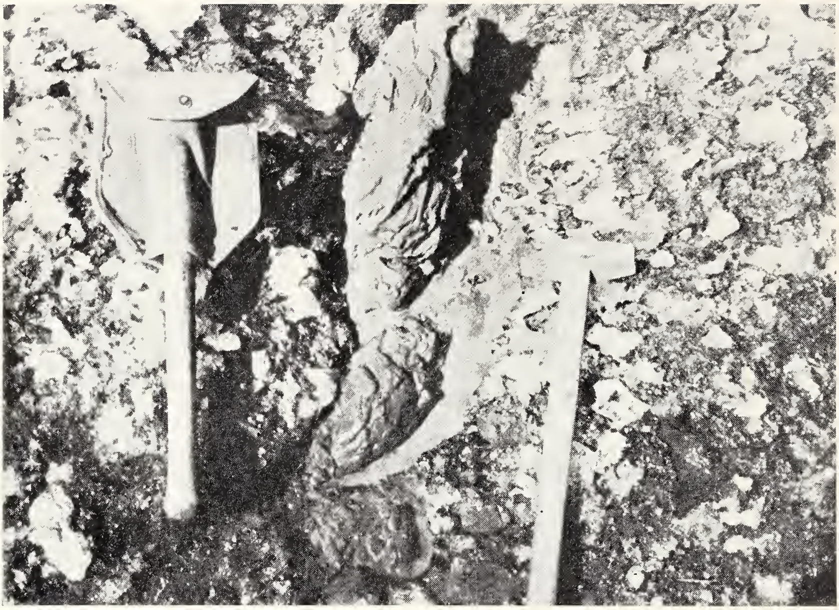

Fig. 7. Upthrust, water-soaked basal load on Hoffelsjökull (between the hammer and the spade).

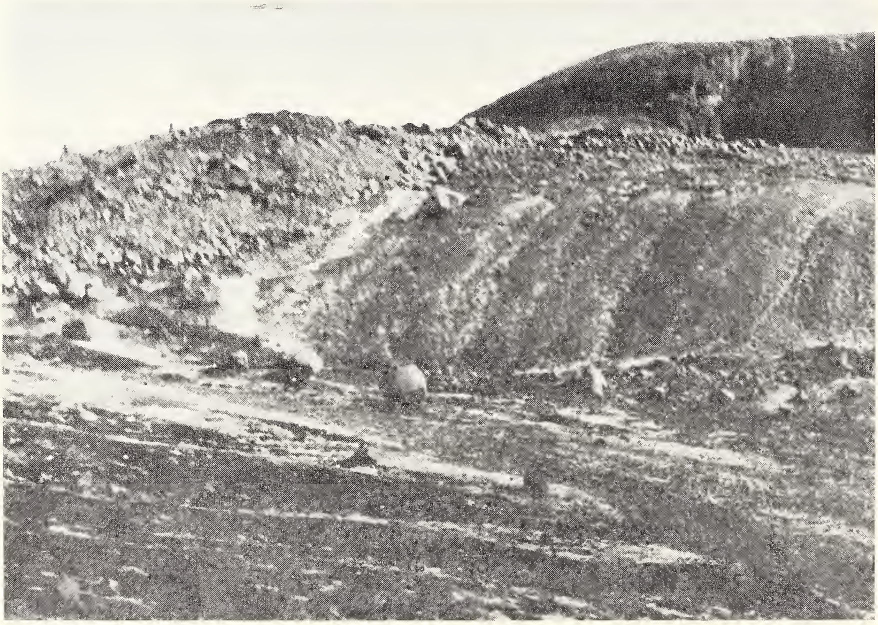

Fig. 8. Drift-covered icewall on the terminus of Hoffelsjökull (left). The dark mountain in the background is Svinahryggur.

PLATE V

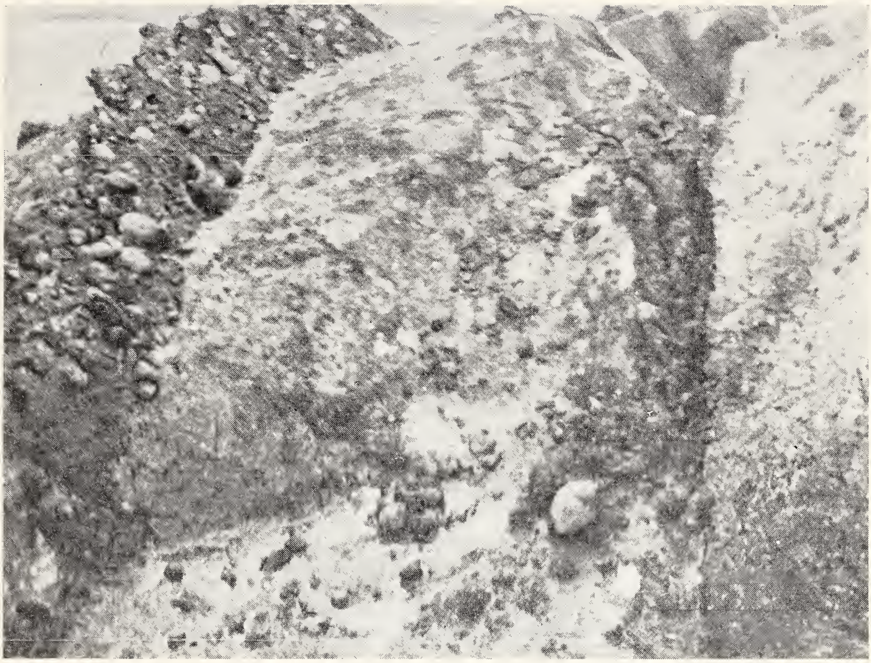

Fig. 9. Transversal or thrust-plane moraine on Fláajökull about 100 m from the terminus (left). Drift appears as a 10 cm thick layer between two sheets of ice and the layer inclines upstream. Owing to the different rates of melting a drift-covered slant plane occurs atop the glacier. The stone in front is about 10 cm in diameter.

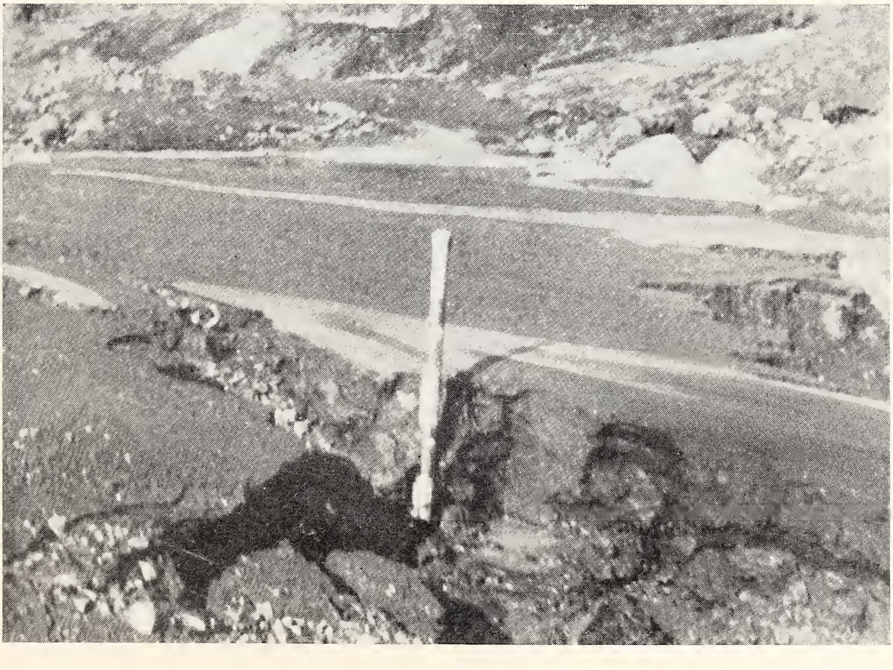

Fig. 10. A small delta on the surface of Hoffelsjökull sloping towards the glacier. The proximal end adjoins the drift-covered icewall seen in Plate IV, Fig. 8.

PLATE VI

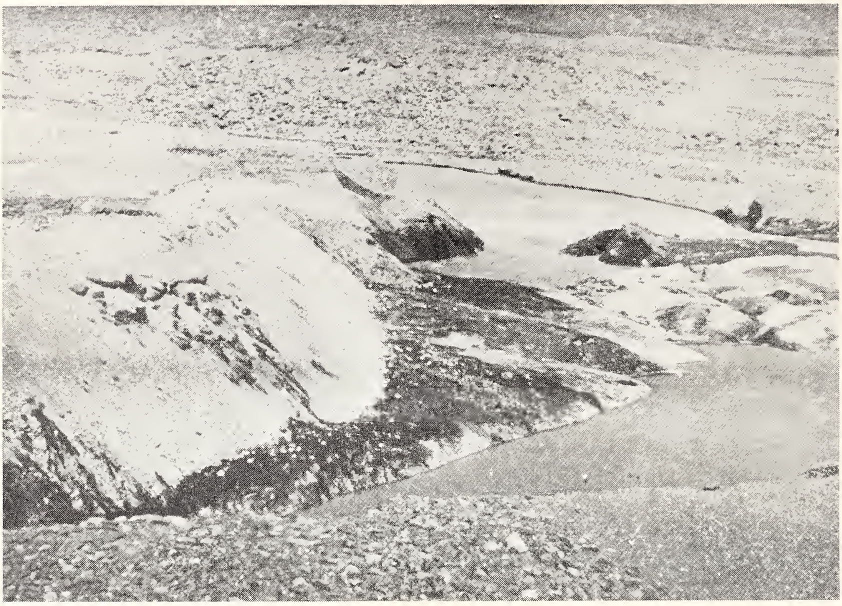

Fig. 11. Laterally flowing stream in front of Breidamerkurjökull cuts its channel through outwash gravel and dark dead ice beneath the drift. The buried ice forms sills in the river bed and continues under the outwash plain.

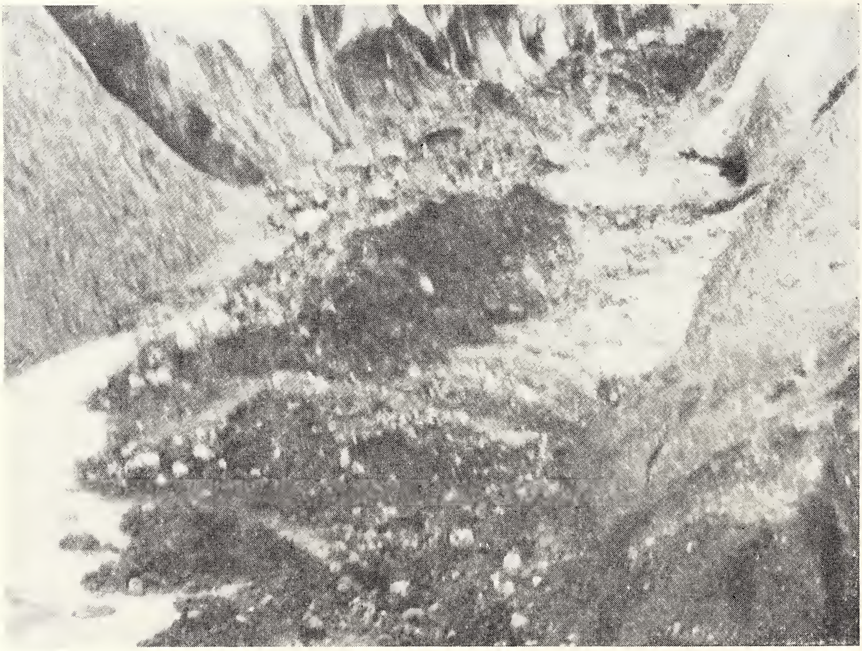

Fig. 12. The bottom of a radial crevasse at Fláajökull. Between and beneath the icewalls there appears water-soaked basal till.

PLATE VII

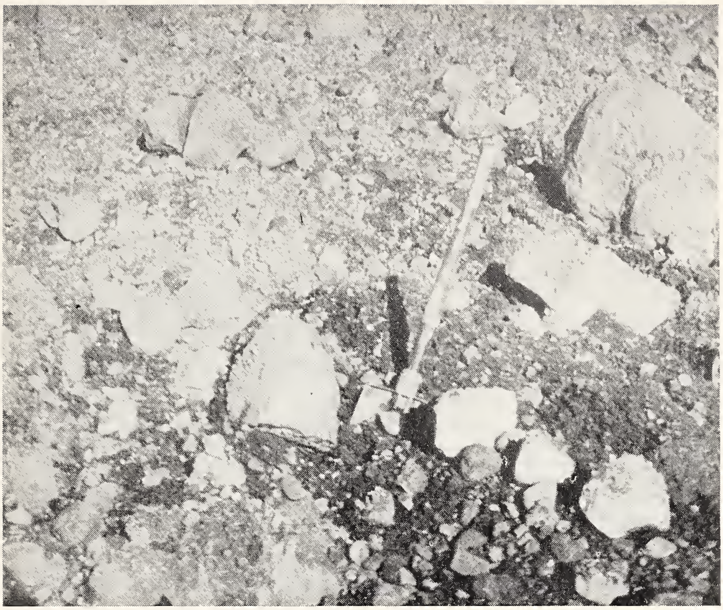

Fig. 13. Ground moraine in front of Breidamerkurjökull is composed of basal till.

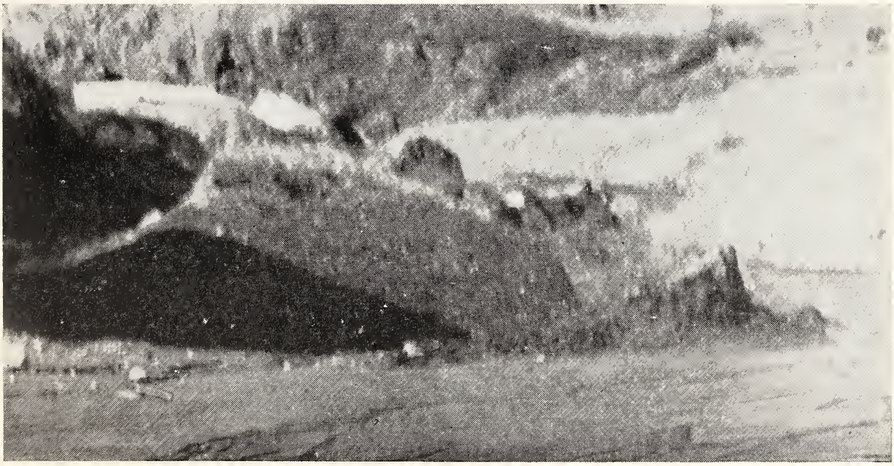

Fig. 14. The projecting eaves of Hoffelsjökull has broken off and dropped into the wet drift pressing it up to crests. These connect with the proximal side of the end moraine (right).

PLATE VIII

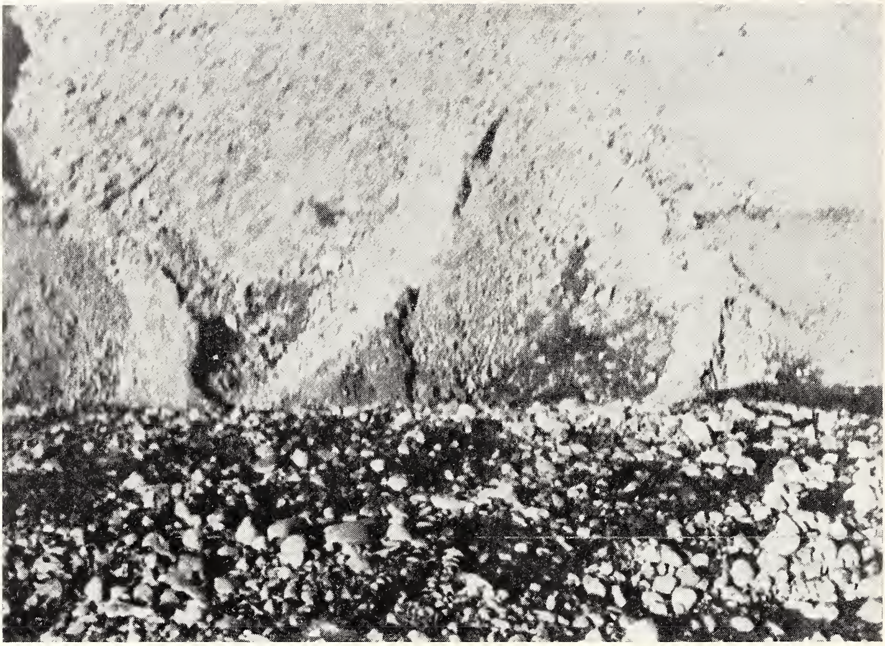

Fig. 15. Small radial moraines seen from the drift-covered terminus of Breidamerkurjökull (foreground). They are imprints of ice blocks that have broken off the glacier snout.

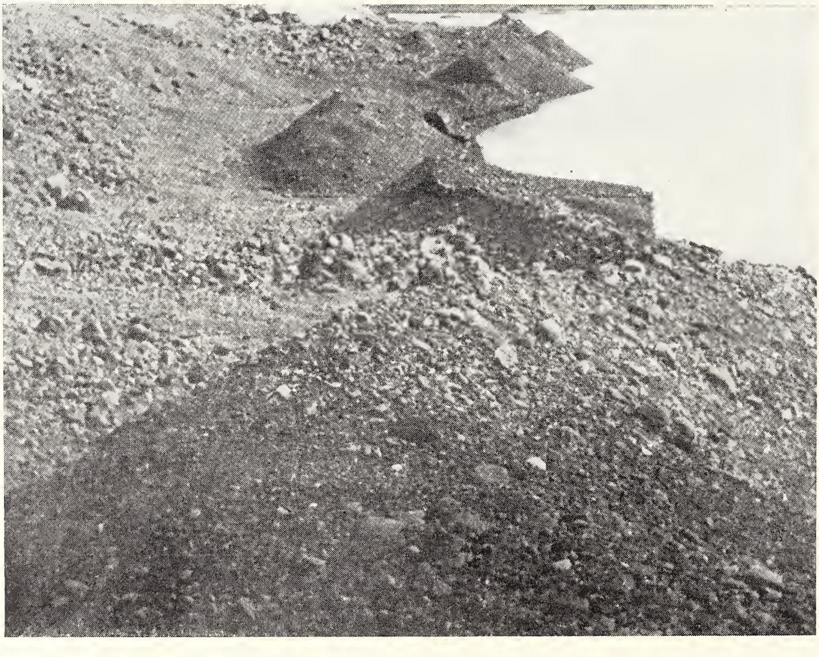

Fig. 16. A radial moraine in the frontal terrain of Svinafellsjökull (west of Hoffelsjökull) resembles an esker ridge along the shore of the marginal lake.

PLATE IX

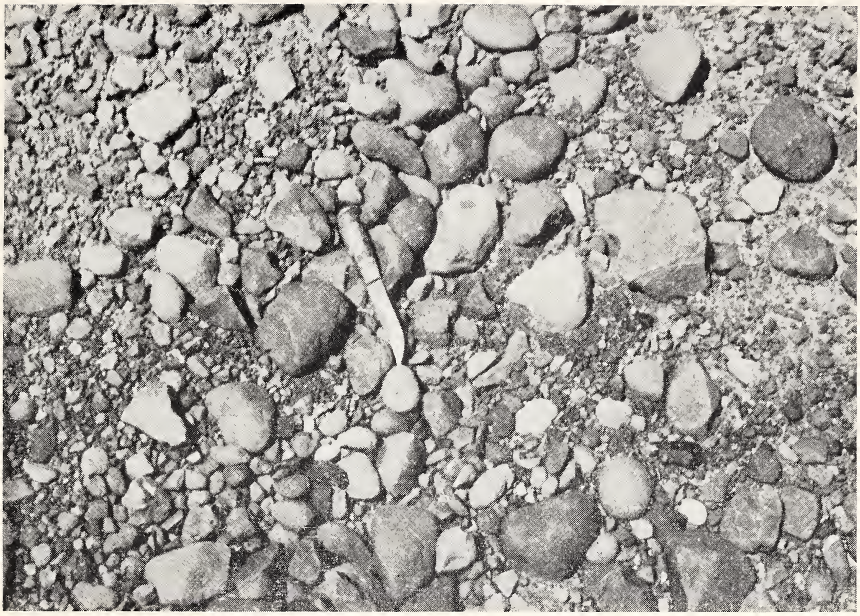

Fig. 17. The material of the radial moraine seen in Plate VIII, Fig. *16* consists of angular glaciated stones and weakly striated roundstones. The length of the knife is 22 cm.

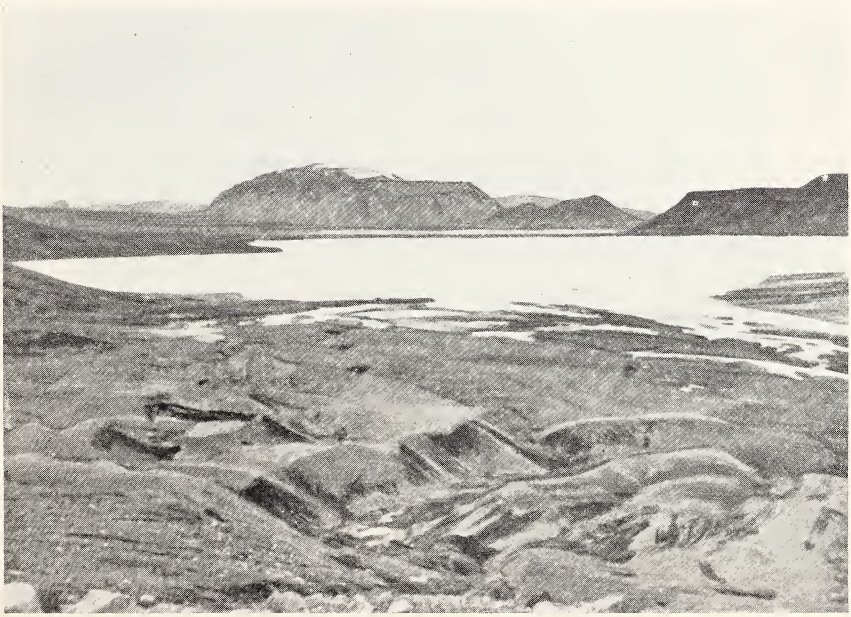

Fig. 18. Forming of ablation moraine on the eastern shore of lake Hagavatn. Black, drift-covered stripes indicate dead ice, which causes sharp forms. The margin of East Hagafellsjökull (right) continues as a thin sheet under the lake level. Photo Helgi Jónasson in 1946.

PLATE X

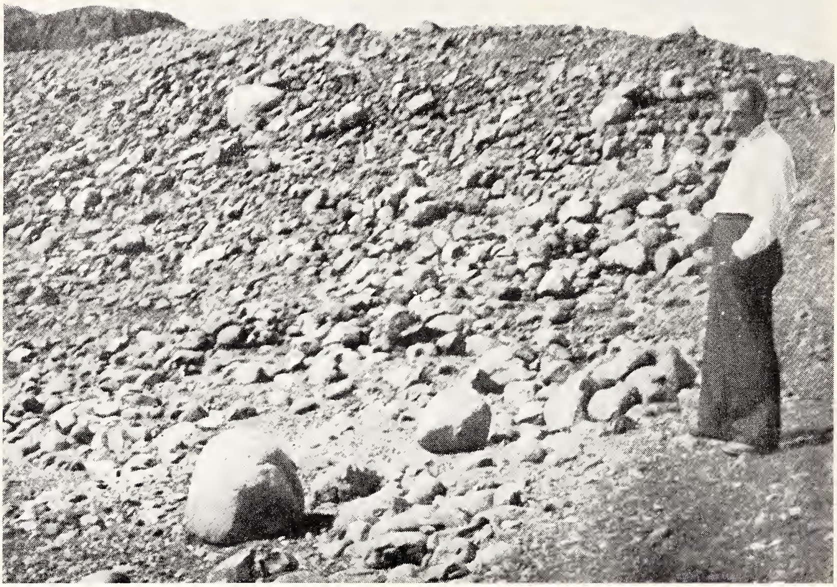

Fig. 19. Surface material on the frontal moraine of Hoffelsjökull consists mainly of roundstones, some of which are weakly striated. After the forming of the moraine (which happened about 1890, according to Thorarinsson) winds have carried the finest particles away from the dried surface.

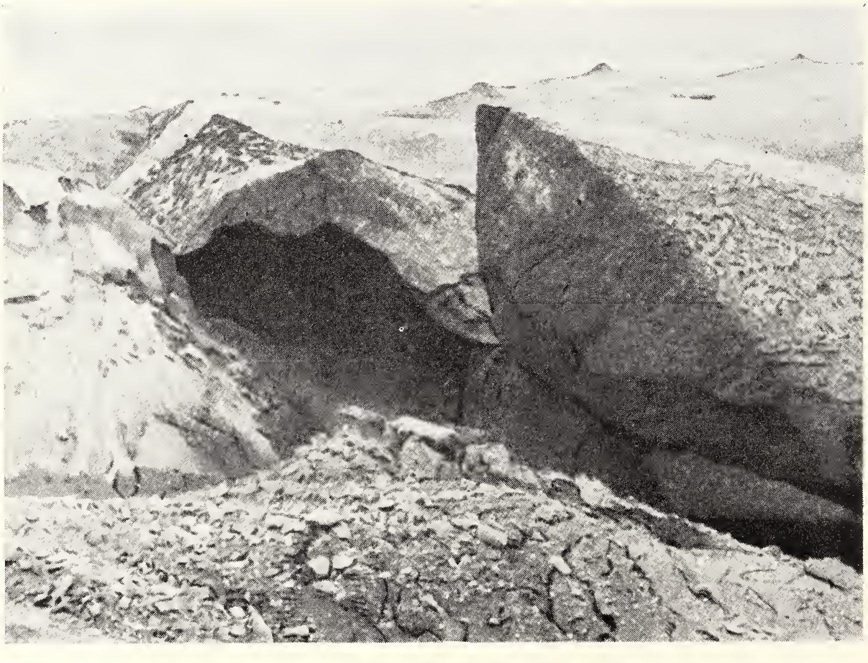

Fig. 20. The terminus of Heinabergsjökull projects vault-like over the marginal pool and basal till. The latter is sliding into the water, and thus becoming trans formed into washed drift.

PLATE XI

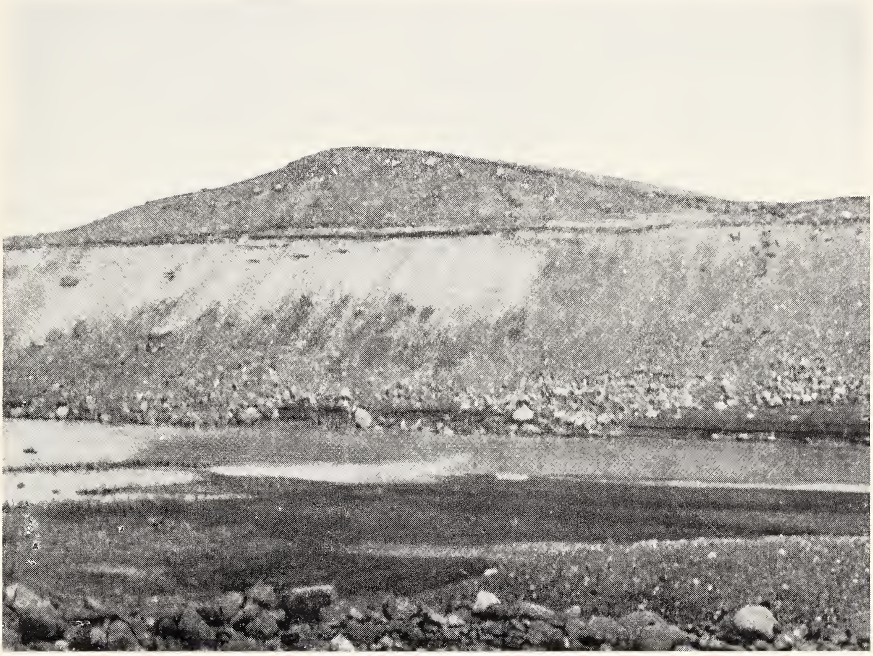

Fig. 21. The bank of the river Hanypá at Heinabergsjökull has been cut through an end moraine and underlying outwash plain. The end moraine ridge is formed on undisturbed sandur deposits.

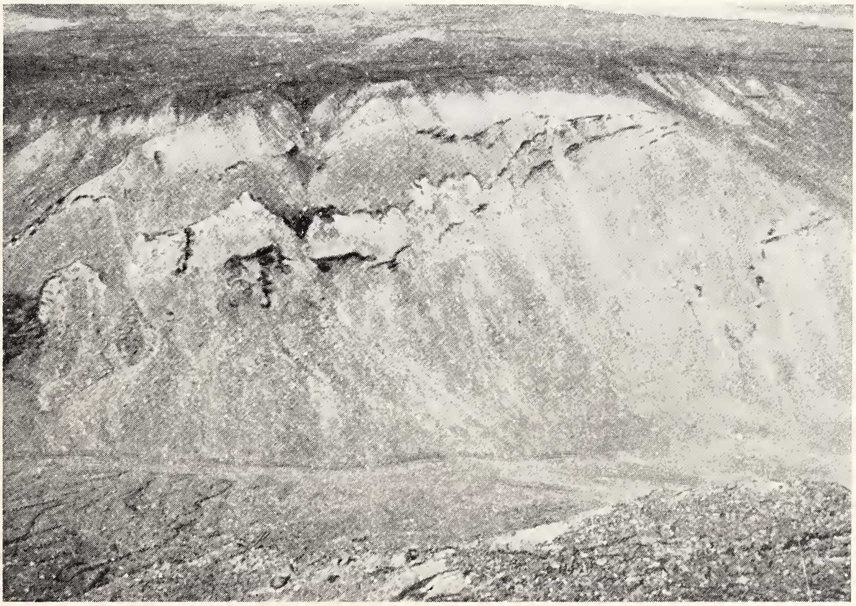

Fig. 22. The vast end moraine belt of Breidamerkurjökull is cut by a dry river bed. In the northern wall there occurs folded layers, which incline to the proximal side (left). The dark bands are thin peat layers.

PLATE XII

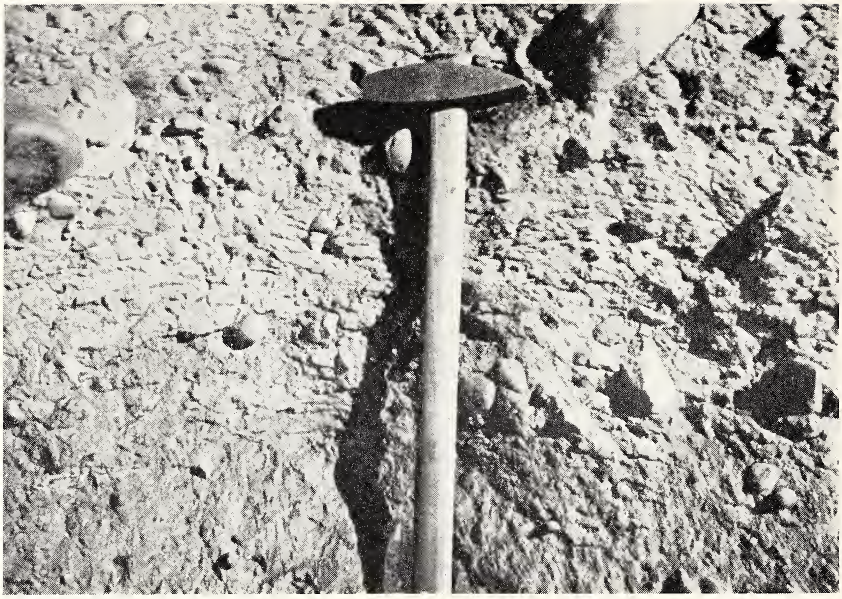

Fig. 23. A bed of compact till covers the top and the proximal side of the end moraine seen in Plate XI, Fig. 22.

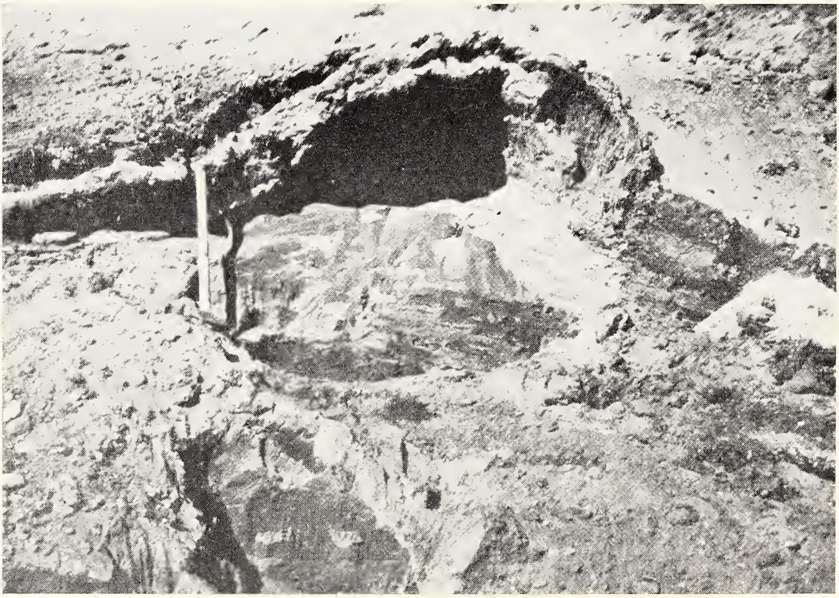

Fig. 24. One of the folded peat layers seen in Plate XI, Fig. 22, and stratified sandy deposit beneath it.

PLATE XIII

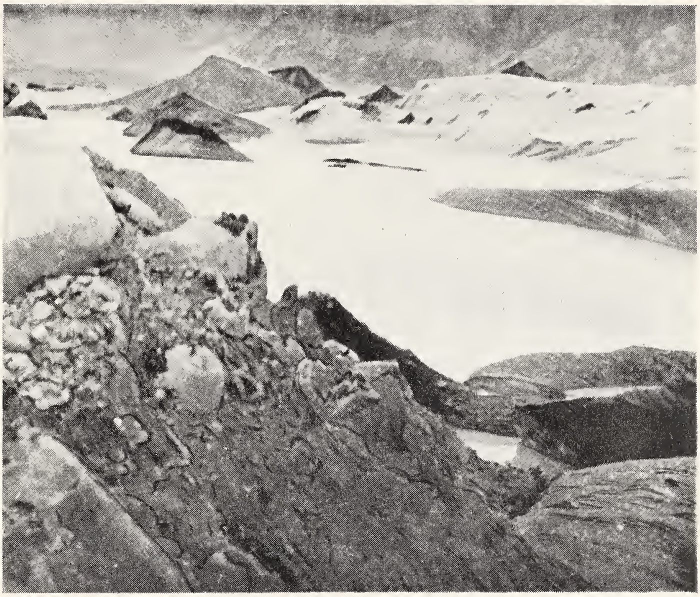

Fig. 25. The Morsárjökull terminates in a marginal lake as a thin subaquatic sheet. The tops of projecting ice islands form planes inclined upstream and their ice is full of thin drift bands dipping in the same direction. The picture is taken from the end moraine.

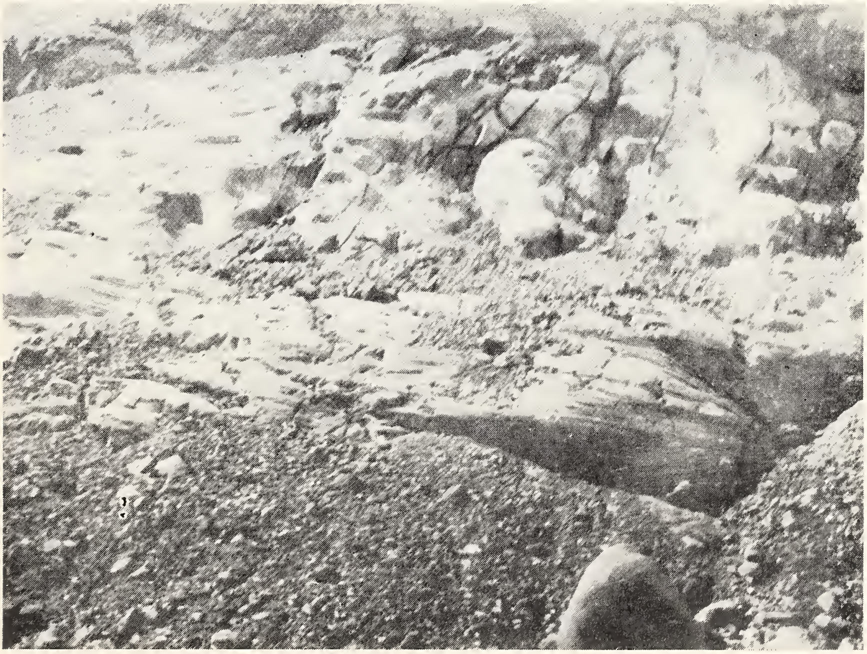

Fig. 26. Rock surfaces on Svinahryggur mountain have a stoss and lee topography with fresh glacial striations in the main ice movement direction. The mantle of basal till is thin.

PLATE XIV

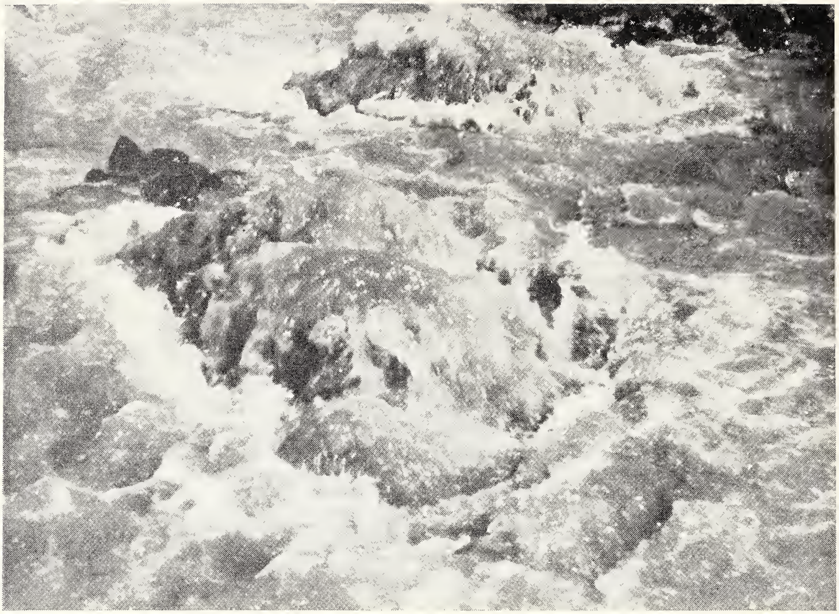

Fig. 27. Artesian wells bring water into the river flowing outside the visible margin of Svinafellsjökull (Hoffelsjökull). The hydrostatic pressure is due to the stagnant ice under the glacial drift. The wells seen in the picture are about one meter in diameter.

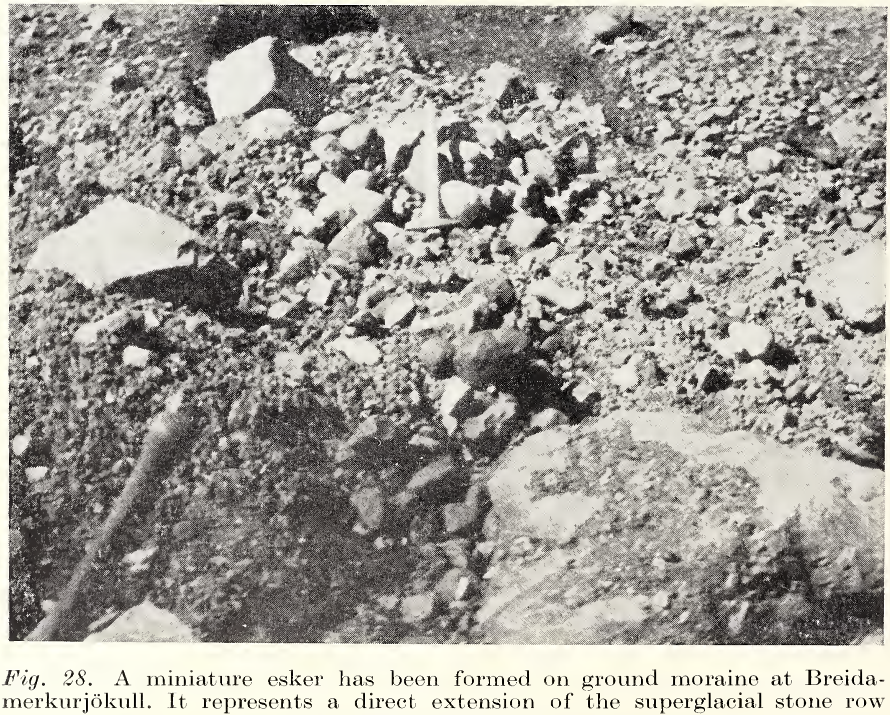

Fig. 28. A miniature esker has been formed on ground moraine at Breidamerkurjökull. It represents a direct extension of the superglacial stone row presented in Fig. 15 p. 38.

PLATE XV

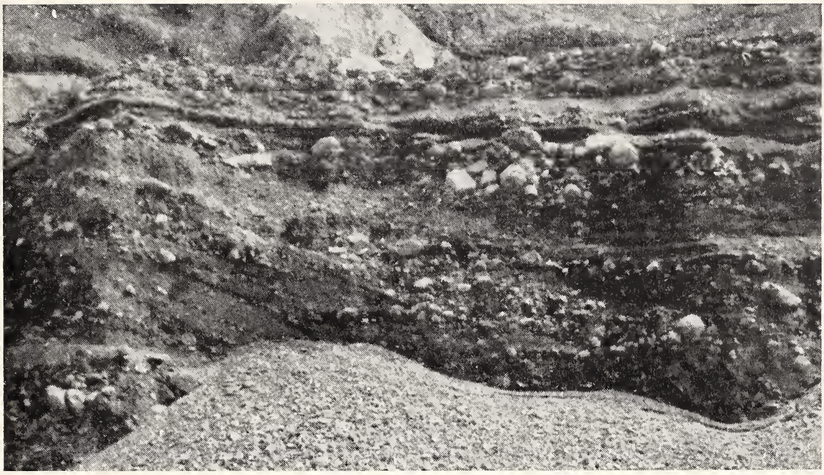

Fig. 29. The bottom sediments of the ice-dammed lake (Efstafellsvatn at Hoffelsjökull) seen in a river bank. The fresh wall in the picture is about 2 m high.

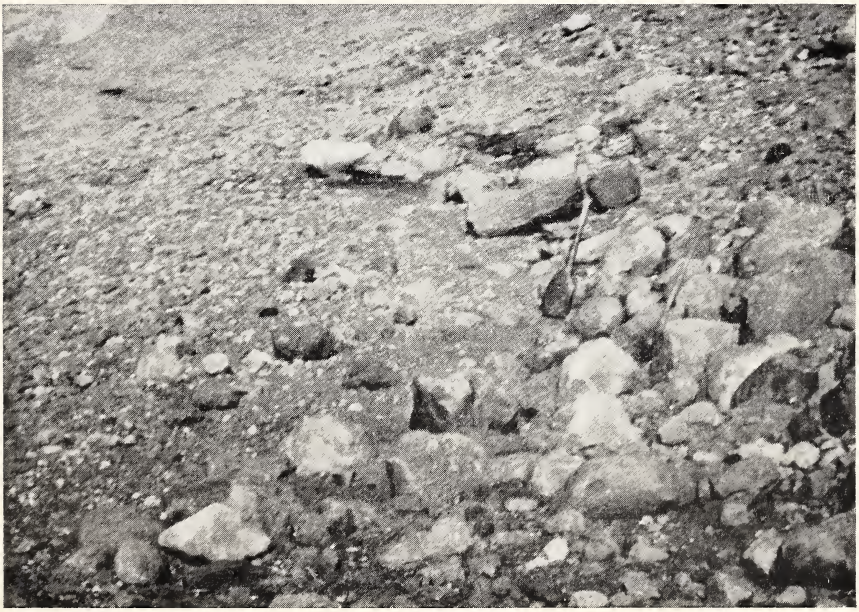

Fig. 30. A typical shore line of an ice-dammed lake (Gjánupsvatn). The spade and a small remnat of a floated ice block indicate the water level before the outburst. The shore becomes more distinct seen from a distance.

PLATE XVI

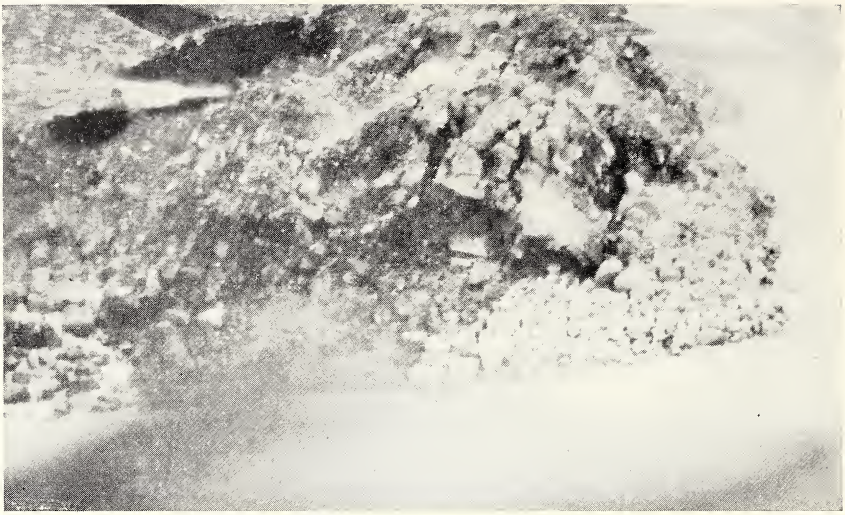

Fig. 31. The border of Fláajökull (top left) covers melted but wet till in immediate contact with water. After being released from the glacier the till is thus reworked, washed and sorted, which changes its character.

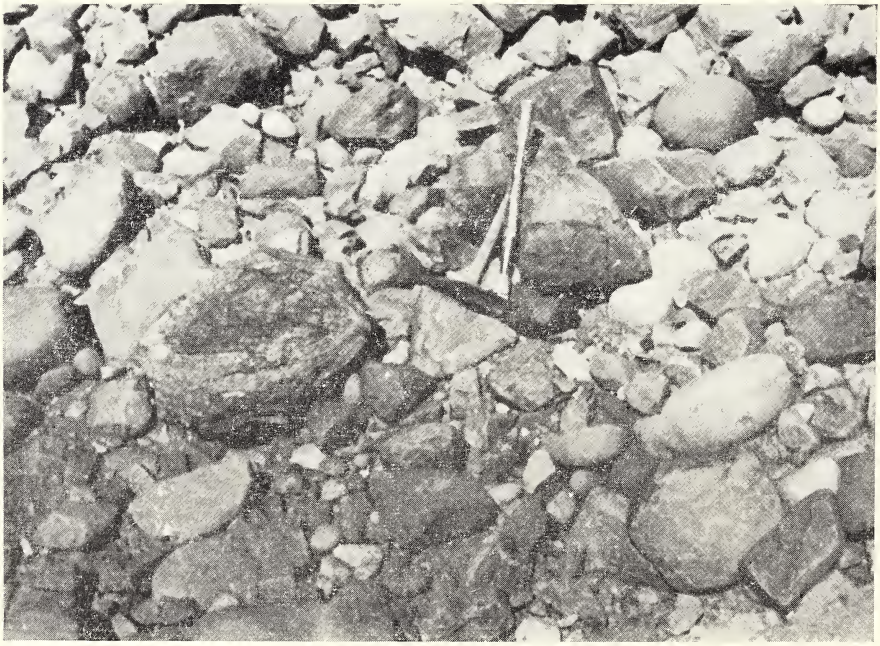

Fig. 32. Coarse material on an outwash plane at Breidamerkurjökull. It has been carried from the glacier by a glacier outburst. Among the stones there appears a chunk of peat (left of the spade).

ACTA GEOGRAPHICA 15, N:o 2

FUERTEVENTURA

SOME GEOLOGICAL AND GEOMORPHOLOGICAL ASPECTS OF THE OLDLAND OF THE CANARIAN ARCHIPELAGO

BY

HANS HAUSEN

HELSINGFORS 1956

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Preface

The island of Fuerteventura, one of the largest of the Canaries, a part of the Eastern Canarian Province of Spain, has up to a quite recent time almost remained outside the tracks of scientific travellers in this insular region. In spite of Fuerteventuras highly interesting geologic conditions geologists have paid but little attention to the island since the German explorer GEORG HARTUNG made his first reconnaissances, including in his study also the neighbour island of Lanzarote. And this journey was made not less than a hundred years ago, under rather primitive circumstances. The results were, however, as we shall find, rather important.

In the year 1950 the author got an occasion to visit the island the first time, in connection with a field work in one of the other islands of the Canaries — Tenerife. That time I stayed here only a week, and I undertook several excursions by car, collecting many rock specimens. A second time I arrived to the island at the end of 1954, and later on in the spring of the following year. Altogether I have spent four months in studies of the geology of Fuerteventura.

The two later journeys were made under the auspices of *El Museo Canario* of Las Palmas and with instructions to perform a geologic reconnaissance survey of the island. The author is highly indebted to *El Museo Canario* and to its *Junta Directiva* for this singular opportunity to investigate this little known region of the Canaries. I am also grateful for the permission obtained to publish this preliminary summary of my results also in a foreign language.

Helsingfors—Brändö, Finland, november 1955 The Author

Introduction

Fuerteventura has a surface extension second in order of all the Canary Islands and comprises 1.731 km². The largest of the islands is as known Tenerife. Fuerteventura lies in the eastern part of the Archipelago and belongs, as mentioned, to the Eastern Canarian Province with its capital Las Palmas of Gran Canaria. Fuerteventura lies not very far from the African coast, the minimum distance being about 115 km.

Fuerteventura has up to the most recent days not been much visited by foreigners, although the ports of the island since a rather long time are in communication with the chief ports of Gran Canaria and Tenerife. That depends of course on the sparse population of the island and its primitive accomodations. Now the conditions have changed to some degree. The capital Puerto de Cabras is provided with a hotel, and the roads have been improved. A regular service by air began some years ago between Puerto de Cabras and Las Palmas, and the air line extends to Arrecife of Lanzarote.

The author performed his survey to a great deal during footexcursions, the only possible means of forthcoming in the open country outside the roads.

Fuerteventura is owing to its desolate nature in the reality a very appropriate field of study to the geologist. The rocks are much exposed, and the dry bottoms of the »rivers» display excellent, fresh exposures of them. The excursions are hampered only by the absence of drink water — in summer time also by the burning sunshine — and by the strong trade winds blowing nearly all the year round.

Collections of rocks and soils made during the excusions, have all been brought to *El Museo Canario,* except what has been kept apart for laboratory treatment. — This kind of work is now going on, and the author hopes in the nearer future to present the results in a more comprehensive memoir.

Earlier observations

The first geologist who visited the island of Fuerteventura was, as pointed out above, Georg Hartung. He accomplished his field work in the year 1854, and its results were published 1857, comprising the descriptive text. The following year a geologic map of Fuerteventura and Lanzarote appeared together with several profiles (comments to them had already been given in connection with the text). We will in the following get occasion to return to this work of Hartung, but we may here in short quote the most important results since they form the basis for all further studies on this interesting island. The rock characteristics are, it is true, superficial; this depends of course on the fact that at the time when Hartung was at work, the science of petrography was still in its »macroscopic» stage; the microscope had just been introduced by Sorby for the study of rocks.

The work of HARTUNG contains three parts: I. Introduction (Physiography), II. The orographic conditions, and III. Geologic conditions. Moreover there are a coloured geologic map of Fuerteventura and Lanzarote together with several geologic profiles. In the geologic part we find at first a short synopsis of the geology of the two islands, then the different lithologic complexes are treated from the youngest to the most ancient. In »Superficial deposits» the limestone cover of the island is described, a characteristic phenomenon. In the sequel the various basaltic series are nearer treated, at first the most recent volcanics (chiefly represented in Lanzarote), then the Sub-Recent lavas, further the old basalts, forming the table mountains in Fuerteventura (as well as in Lanzarote). The last chapter deals with the so called »Syenite- and Trapp-formation» represented only in Fuerteventua. Finally the work con-

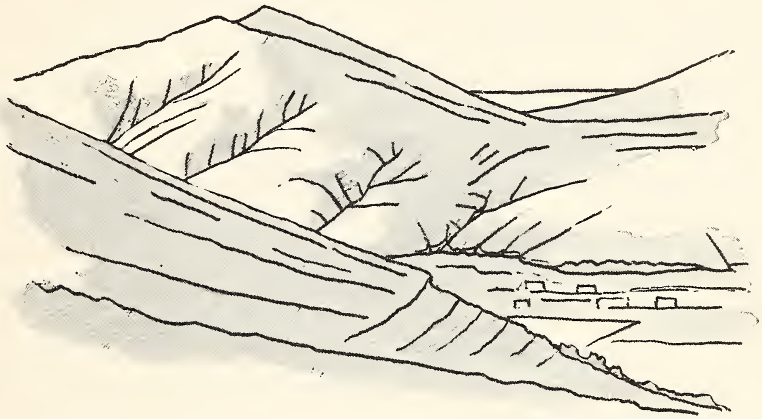

Fig. 1. Montaña de Atalaya (726 m) the highest top in the Betancuria Mountains. Looking west. The village Betancuria in the valley bottom.

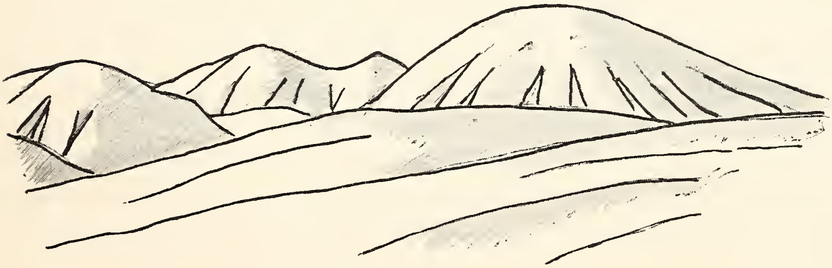

Fig. 2. Montaña de la Blanca (308 m), to the west from Oliva, consisting of gabbro- and perioditic rocks (covered by travertine, hence the name). Looking northwest.

tains some concluding remarks. Moreover the geologic profiles are accompanied by comments.

No doubt the chief interest is offered by the chapter on the »Syenite- and Trapp-formation», exposed in the broad mountainous zone accompanying the northwestern coast. HARTUNG came to the conclusion that the deep seated rock-masses (»Syenite» etc.) appearing in the central part of the mentioned mountains are the oldest basement of the island, and that this core is capped by a series of basalt lavas called by him »Trapps», a volcanic formation of more ancient date than all the other basaltic lavas on the island. There exists a marked unconformity between the Trapps, standing on edge, and the overlying (now mostly denuded) basalt series of a table-land formation. Dikes in the Trapps are common, also in the plutonics, Many of these dikes seem to be trachytic in composition.

HARTUNG is of the opinion that the mentioned oldest structural element of Fuerteventura may belong to the same stratigraphic unit as the similar rocks exposed in the deepest parts of Gran Canaria and La Palma on which all the younger lavas have been piled up.

Somewhat later (1863) KARL VON FRITSCH visited Fuerteventura during his long campaign in the Canaries. Tenerife was thoroughly investigated by him (together with W. REISS). Fuerteventura was as it seems more rapidly studied. In his paper »Reisebilder von den Canarischen Inseln» (1867) he gives also a good picture of Fuerteventura, its nature and inhabitants.

In the time to follow the Spanish geologist S. CALDERÓN Y ARANA of El Museo Canario in Las Palmas made journeys to the different islands, and he visited also Fuerteventura. Of his reports only one treats Fuerteventura in a study of the sands and dunes of the islands (1884). In this sense Fuerteventura is of mayor importance.

In the course of several decades no more geologists found their way to Fuerteventura. We have only to observe a small publication on the Fuerteventura rocks by L. FINCKH (1908) treating the plutonic as well as the dike rocks appearing in the mountains around Santa Maria de Betancuria. FINCKH was never here, he only studied collections brought to Halle (Germany) by K. VON FRITSCH and guarded there. He found that the plutonites appearing in the mountains in the vicinity of Betancuria and Vega Rio las Palmas (mountains bordering the northwestern coast) are intrusive in the series of »Trapps». These deep seated rocks stand in connection with many dikes of similar composition.

In the second and third decade of the present century Fuerteventura was repeatedly visited by members of the Engineering Staff of Cabildo Insular de Las Palmas in connection with different public works in the island. S. BENITEZ PADILLA of that department, an ardent student of Canarian geology, made during these journeys several observations. They were not published, only a theoretic synopsis of the Canaries including Fuerteventura has lately appeared (1945).

In the year 1934 Fuerteventura was visited by the well known French geologist JACQUES BOURCART with the intention to study the ground water conditions and the general geology, commissioned by Cabildo Insular de Las Palmas. He made a somewhat rapid reconnaissance collecting much material among the rocks of the island. After the laboratory work was ended he published together with E. JÉRÉMINE a geologic report on the island (1938), containing many petrographic characteristics and also chemical analyses of the leading rock types.

Physiography

Fuerteventura has compared with the westerly islands of the Canaries a much more open relief, although there are plenty of mountains. But the valleys are generally open with broad, flat bottoms, and wide basins are likewise present. The coasts are to a great deal low and sandy, in some sectors steep cliffs are, however, present. The elevations do not attain great measures; the highest point is Pico de Jandía 807 m, next in order there is Montaña de Atalaya in the region of Betancuria with 726 m. (Fig. 1). Other hills and ridges attain about 400—500 m including several volcanic cones, of which most are lower.

Orographically the island can be divided ino the following parts: 1. The western Cordillera or Montañas de Betancuria, following the northwestern coast as a broad zone, measuring about 40 km in length and 20 km in width in the broadest sector. 2. The longitudinal valley, in the northern and the southern ends widened to plains (basins). This valley has a broad treshold in the center of the island. 3. The eastern Cordillera, a row of more or less isolated, transversely oriented ridges, following the southeastern and eastern coast. 4. The peninsula of Jandía, a rough mountain ridge, separated from the rest of the island by the low Istmo de la Pared. 5. Groups of volcanic cones and also isolated ones scattered over the island surface.

All the coasts are open; no natural harbours are present, only embayments. Sheltered places can be found here and there along the leeward side of the island (the »*sotavento*» coast). The total length of the coasts measures up to 265 km.

Most of the ground is barren and desolate, and the effects of the weathering can be seen everywhere. Work of the running water is likewise apparent in spite of the extreme dryness that dominates the island. The erosion is put into action only when some cloudbursts occur, what happens rarely in the course of a whole year (winter season).

As we shall find later on the whole physiognomy of the island is nevertheless dominated by the erosion forms and the sediments consisting of the detracted material. We have to conclude that there were different climatic conditions in ancient times.

Owing to the dryness of the climate [1] the population is sparsely distributed

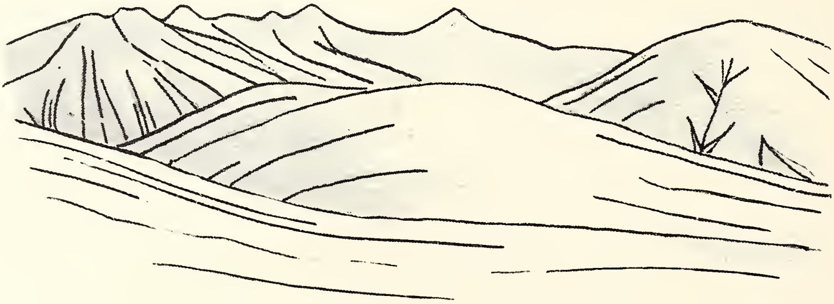

Fig. 3. The southern declivities of Montaña de Ancones, consiting of the Trapp-formation, Looking north-northeast from a foothill. Highest elevation of the ridge: 606 m.

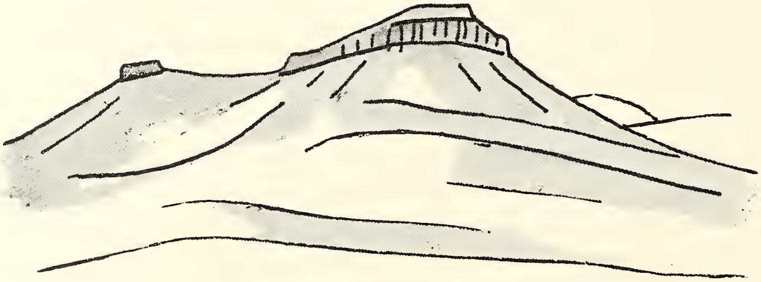

Fig. 4. An erosion hill at the southern edge of the Tuineje basin, consisting of the basal breccia formation of the table-land basaltic series capped with a hard sheet of the last named. Looking south.

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[1] About 200 mm of rain annually.

and attains hardly more than 15 000 souls. Cultures of maiz, potatoes and alfalfa have been made possible only by irrigation with ground water in the lowlands. Tomatoes are here and there also cultivated though still on a small scale. Capital of the island is Puerto de Cabras on the eastern coast. From here one highroad runs to the north — to Oliva and another southern branch to Gran Tarajal, passing through the old center of the agriculture region — Antigua. Also the ancient capital of the island — Santa Maria de Betancuria has been connected with the highroad by a recently constructed road running through Pájara.

Geologic summary

Before entering into the descriptions of the geologic development of the island we may in general terms mention the present status of the knowledge of the geologic structure.

The Canary Islands are as is known chiefly of volcanic nature. Especially the western islands including Gran Canaria represent imposing accumulations of volcanic material attaining heights of 2 000—4 000 meters above the ocean level. And the submarine parts of the islands in question may measure heights of 2 000—3 000 meters. Looking at Fuerteventura we will find the things rather different, and a great difference can be noted also when comparing it with the northern island Lanzarote, the volcanic island *par préference.*

Fuerteventura exposes over wide areas an old ground consisting of a mighty succession of lavas upturned on edge — a folded zone as it may be characterized. These old lavas have been intruded with deep seated magmas of varying composition, and a multitude of dikes has penetrated the formation, mostly in the longitudinal direction, which is NNE—SSW. The plutonic bosses have caused a kind of contact-metamorphism in the surroundings.

Over the upturned edges of the old lava formation called »Trapps» by G. Hartung there lies in a rather flat position a series of basaltic lavas, tuffs and agglomerates (with a conglomerate formation at the base) attaining in some cases a thickness of several hundred meters. This formation is slightly inclined towards the eastern and southeastern coasts. It has been strongly denuded, so that only remnants are to be seen (composing eastern Cordillera).

Next in age are some broad shield-volcanoes of basaltic lavas appearing in the valleys and filling with their numerous lava streams a great deal of the bottoms.

The time separating the old table-land series from the last mentioned lava volcanoes was one of faultings and erosion. Also huge masses of detrital material were accumulated in the valleys before new eruptions took place.

Later on a volcanic cycle was initiated producing a great number of cinder cones, the forms of which are still rather well conserved. This activity began at the final stage of the Tertiary era and continued into the Quaternary period, perhaps to a Sub-Recent time, although with intervals of different length.

Cones belonging to the »historic time» (in European sense) have not appeared in Fuerteventura. LEOPOLD VON BUCH who states that the volcanoes in the north of the island were active in the XVIII:th century is in error. The volcanic catastrophes devastating the island of Lanzarote at that time did not reach to Fuerteventura, not even to the island of Lobos in the Bocaina Strait.

Jandía, the strange appendix at the southern end of the bulk of the island has been interpreted as the only visible segment of a giant caldera of basaltic lavas, now to a mayor part conquered by the sea. The lavas may belong to the old table-land series.

The volcanoes of Fuerteventura represent a very distinct constructive morphologic element contrasting against the destructive erosion forms of the mountains in the west as well as in the east.

Aside from the volcanic formations with their lavas, tuffs, agglomerates, and cross-cutting dikes there are also some sedimentary formations to be observed. At many points along the coast remnants of a marine foraminiferan firmly diagenetized limestone rest on abrasion terraces. The lime ooze was deposited when the island in relation to the ocean surface lay several hundred meters lower than now.

Further there are deposits of different kinds in the valleys and basins: tuffs and lapillis more or less weathered and decomposed, clays, sands and gravels of fluviatile origin. Of great importance is the travertine cover extending over practically the whole island, including the mountain slopes and reaching right down to the sea border. This cover of calcium carbonate is purest at the surface, but downward it is more and more mingled with lava fragments loosened from the underlying ground. No one of the Canaries has such a mighty plate of travertine limestone as Fuerteventura.

Another sedimentary deposit is represented by a calcareous sandstone of brittle coherence, showing cross bedding and a rusty yellowish colour. It is a terrestric formation mostly destroyed by later erosion. It contains also terrestric fossils and petrified nests of the genus *Antophora*. This sandstone has furnished copious masses of drifting sand in some parts of the island (for instance on the NE coast and Istmo de La Pared).

The distribution of the different formations has been fixed approximately on the sketch-map of the island, plate I, Fig. 1.

The geologic development

1. The ancient volcanic series

The author has devoted much time to excursions in the many valleys and gorges that dissect the western mountain block in different directions. My studies have comprised practically all the sectors of this region from Montaña de la Blanca in the north (region W from Oliva), to the surroundings of Cortijo de Chilegua in the south. The most impressive behaviour of the mountain structure is its upturned series of basic lavas (»Trapps») and their very constant NNE strike direction. There is a mountain axis following the inner structure, and this is well exposed on a distance of about 40 km. Another striking fact is the steepness of the dip of the lava beds (with their parallel dikes); either vertical or tilted under 80°—70° to the WNW or ESE. Another remarkable fact is the appearance of deep-seated plutonic bosses in the center of the area, known since the time of G. Hartung (1857).

We find in this region a part of an ancient fold-zone with tightly compressed isoclinal sheets of lavas invaded by deep seated magmas and their dike protuberances. In the extreme north the fold-zone in question disappears under young volcanic accumulations, in the extreme south (SSW) one sees how the upturned edges of the lavas submerge into the ocean (see Plate III, Fig. 5), a locality just to the north from Playa Ugán.

There are inside the mountain region plenty of occasions to study the details of this fold-structure. Since the material collected is still under laboratory research the author does not enter into petrographic characteristics, but will only make some general remarks.

The upturned beds of lavas are mostly fine-grained and dark greenish in colour. Jérémine has (1938) identified them with spilites, and they are always altered. Finckh (1908) called lavas occurring in this complex primary basanites and trachydolerites, which later on were altered as a consequence of plutonic intrusions. The author has not yet studied the types more closely. — Apart from the dark greenish trapp-lavas there are perfectly concordant sheets of pale-grayish — greenish colour, rocks that may be very rich in feldspar and reveal a bostonitic texture. They seem to be mostly altered and can be classified as primary trachytes. There is a vivid alternance between these two kinds of lava rocks, of which the latter may be interpreted as dikes (?). Their relative quantity in the complex seems, however, to be astonishingly great. Black basaltic dikes are also present, and they run generally concordantly with the other members of the complex (camptonites, according to FINCKH).

There is no possibility to estimate the total thickness, since the whole mass shows a perfect isoclinal fold structure. The maximum transversal dimension of this zone is about 20 000 m, but there may be many recurrences. The formation was once extended over the earths surface in a flat lying position, and the lava was certainly very fluid. Tuffs may have been intercalated. Such ones are mentioned by E. JÉRÉMINE (1938).

2. The upfolding of the ancient volcanics whith the intrusion of bosses and dikes

We have found from the foregoing chapter that the old volcanics are strongly compressed to a concordant pile of lava sheets standing on edge, either in a vertical position or inclined 80°—70° to one side or another of the axis of deformation, which runs SSW—NNE (with small deviations). This remarkable persistency in the strike and also in the dips should be inexplicable if we supposed that the deformation in question would be a more local phenomenon, especially when we take into consideration the intrusive bosses present here. We have consequently to suppose that the fold structure in question may be only a short section of a long fold-zone, whose continuation is to be found in the bottom of the ocean in the general strike direction. There would be expected exposures of the zone in Lanzarote, but here the later volcanic accumulations have covered the basement completely.

The compression of the lava series, which resulted in a steeply dipping position of the complex, leads one to think about some geosynclinal »root zone». There are, it is true, in the upfolded series few signs of deepseated metamorphism (schistosity and recrystallization), but instead more superficial alterations (spilitization etc.). The presence of intrusive plutonic rocks intermingled with the fold complex together with contact-metamorphism and a kind of (more local) migmatization observed by the author speak in favour of a part of an orogenic zone exposed by denudation.

The author has observed on his routes throughout nearly the whole area only a steep inclination of the sheets of lavas (and dikes). Hence I have no idea about the position of the front of the fold-zone and the direction of the compressing power.

According to observations made by G. HARTUNG (1857) the series in question seems to change its position in the direction toward the western coast: the inclination (the dip) of the sheets gets more and more gentle and attains finally only 3°—2°. Personally I have not been able to confirm this statement. During excursions to the vicinity of the west coast I have always found the sheets standing in their usual steep position.

The deep-seated intrusive granular rocks called by HARTUNG summarily »syenite» have been more closely studied by L. FINCKH (1908), and he found a series of Na-rich more salic types such as nordmarkites, pulaskites, akerites etc. Moreover the collection studied contained essexites and dike rocks (camptonites, gauteites).

E. JÉRÉMINE (1938) has found syenites, partly hyperalcaline, diorites, gabbros and ultrabasic types. If all these types are consanguineous or not could in the field not be decided. The microscopic study speaks in favour of the former alternative.

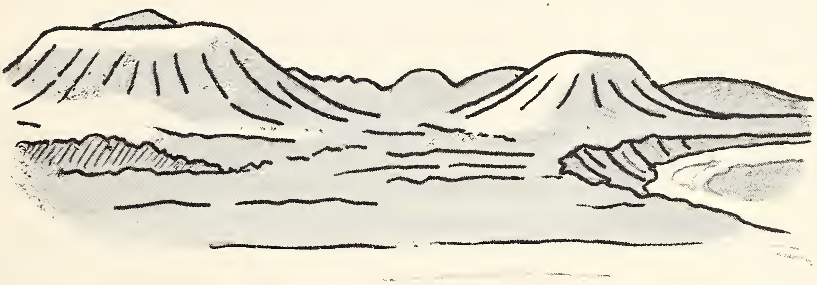

Fig. 5. Montaña de la Mancha (151 m) and a somewhat lower neighbour (to the right) at the northern coast, consist of a brown tuff formation covered with basalt. Looking southwest.

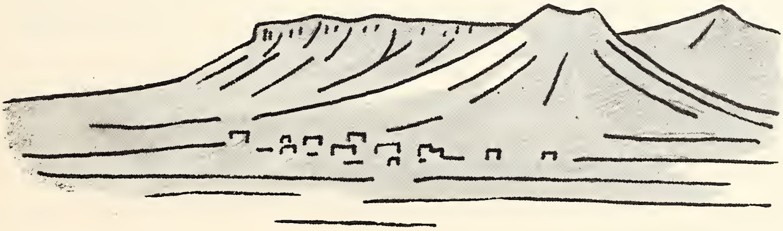

Fig. 6. Castillejo Grande (494 m) of the basaltic table-land series, forelaid by a cinder cone, La Caldereta (392 m). Region of Tetir. Looking south.

The author has collected a fairly representative suite of these granular rocks and tried to fix their exposures in the topographic map. Since the laboratory examination is still unfinished, details will not be given here. It may only be mentioned that the gabbros and their ultrabasic derivatives have been found by me also farther to the north, in the region of Barranco de los Molinos and in Montaña de la Blanca (fig. 2) W from Oliva.

The plutonic bosses of salic rocks are of insignificant extension as was pointed out already by earlier explorers.

3. Subsequent denudation

The fold-mountain range existing in this part of the Canarian area comprising probably also the present Lanzarote and Gran Canaria was in all probability of a bold relief and hence strongly attacked by the weathering and denudation. The crests were continuously worn down, the material loosened by weathering was carried down the slopes and further to surrounding lowlands, where it came to rest. The denudation progressed, as it seems, to the final stage of degradation, — a peneplane was finally worked out truncating all the lava sheets in their upturned position. The cut represented by this peneplane reached such levels that the plutonic bosses were exposed to the daylight and likewise their metamorphic aureoles (E. JÉRÉMINE 1938).

The author being aware of the existence of the deeply reaching denudation surface did have his attention directed to possible occurrences of detrital material deposited in connection with this denudation process. Such remnants may be present somewhere in the surroundings, I therefore examined the fault scarps facing in the east the longitudinal valley and belonging to the eastern Cordillera. Here I found in some places the formation sought for: underlying the basaltic table-land series a coarse conglomerate material in which blocks and stones of all sizes are mingled without any apparent stratification. The best exposures are in Montaña del Campo and de la Fortaleza (fig. 7), as well as in Montaña del Tao lying between them. Here the detrital formation appears with at least 100 m thickness. This conglomerate continues probably to the east, since loose boulders of plutonic rocks were found at the foot of Montaña Enmedio lying between Casillas del Angel and Puerto de Cabras.

Another locality where the coarse conglomerate was encountered, lies close to the *barlovento* coast near to the mouth of Barranco de la Cruz, one km to the south from Playa Molinos. It is a small remnant adherent to the underlying upturned edges of the Trapp-formation. Likewise of insignificant extension is another seen in some exposures along Barranco Largo in the midst of the southern basin, quite distant from the mountain edge (Los Ancones, fig. 3). The basement consists also here of the Trapp-formation.

The scattered occurrences of a coarse detrital material enumerated above enclosing also huge blocks, — all consisting of rocks met with in the mountain chain —, are the witnesses of a transport using gorges and valleys leading down from the fold-zone in question. That there are no more conglomerate masses exposed in the surface of the surroundings, depends on the relatively deep position of the formation and on the covering table land series, to which we will turn our attention in the following. In earlier times there was certainly much more of the formation to be seen, but later erosion has destroyed this relatively soft material. It is possible that the underground of the longitudinal valley still contains mighty masses of the same conglomerate formation owing to the downfaulted position which guarded it from a too rapid destruction.

Hence we have found that the oldland corresponding to the present Fuerteventura and its surroundings (now lying under the ocean), which appeared as a lofty mountain chain, in the course of time was worn down to a peneplane, partly barren, partly covered with detrital material. The latter was getting mightier in the more distant surroundings (the foreland). It was an extremely monotonous land surface, a shield, and it was possibly united with the near lying African mainland.

If we look at the other members in the Canarian Archipelago, we will find similar old denudation surfaces at least in some of them. Studying the deep Caldera de Tejeda in the center of Gran Canaria the author has found a stratigraphic break of a somewhat identic nature. Here a series starting with basaltic lavas has capped a basement — deeply disclosed by the gorge of Rio de Tejeda — which consists of strongly inclined salic lavas intruded with masses of granular plutonics, especially syenites. — In the center of La Palma (Caldera de Taburiente) we will find conditions of the same nature: an old basement with plutonic intrusions mingled with steeply inclined old lavas [1] and capped with a series of younger basaltic lavas in a flat lying position. As far as Tenerife is concerned the erosion has not been able to prepare any deep insights into the interior of the island. Only a multitude of ejecta consisting of »sanidinites» (syenites) lie in the surroundings of the great (central) Caldera de las Cañadas, These boulders apparently derived from the deep interior of the island.

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[1] VON BUCH (1825) identified these with metamorphic shists.

4. Areal effusions of basaltic lavas. The table-land and its denudation

After a period of peneplanation of the old fold structure in these parts of the Canaries, when also masses of detrital products were accumulated in the surroundings of the fold-mountain chain, the crustal quiescense was interrupted by earth movements along fissures, and the major of these cracks opened the way from the earths interior for very fluid basaltic lavas. One such line of eruptions was as it seems situated off the present *barlovento* coast of Fuerteventura. Linear eruptions took place expanding over the surfaces in the east. A remarkable center of eruptions was situated more in the southwest, a part now inundated by the sea. A giant caldera was formed here, but of this volcanic edifice only a segment has been left behind — the Jandía Peninsula.

The flows of basaltic lavas emitted from all the vents were copious and of remarkable uniformity, regarding the composition of the lavas as far as is known. The effusions were preceded by a kind of agglomeratic avalanches, expanding over vast surfaces — or nearly over the entire island, and they deposited a »breccia»-like mass that now forms the base of the table-land lavas. This agglomeratic or »breccia» rock is when weathered, of a soft consistency.

The flows of lavas were numerous and apparently concordant. Some of the lava sheets are of great thickness. Many dikes intersect the series, mostly of basaltic nature, but there are also trachytic dikes and associated with them small laccolithic sheets of pale-coloured trachyte intercalated between the dull-coloured basaltic lavas. A remarkable standing »plug» of trachytic composition is the conic Montaña Tindaya in the northern part of the island. This very isolated mountain has been laid free from the enclosing basalts by erosion (see fig. 11). The mentioned trachytes may be regarded as salic differentiates to the basaltic magma, as is the case with the similar plugs met with in Tenerife.

The old table-land was, after the last effusions had ceased and consolidated, subjected to weathering and erosion. A drainage pattern was developed with eastwards and southeastwards directed gorges. From the coasts in the surroundings the sea attacked the lavas, and the circumference was steadily reduced. With progressing erosion in the land surface the gorges were widened to valleys, and the upper parts of the table-land were denuded.

Of this old drainage pattern we will find a great deal conserved in the eastern Cordillera: in the shape of cross going valleys. In the transverse profile most of them are broadly U — shaped; the bottom grades gently towards the sea. The valleys are directed partly to the east, partly to the southeast, and — at the southern coast — to the south; hence they are oriented perpendicularly to the existing coast-line. All the valleys point to an origin in the western highland at the *barlovento* coast and to its northern continuation (where the highland has been lowered to a series of rounded hills).

Curiously enough the valleys in question — lying transversally to the general trend of the eastern Cordillera — have no begin and no mouth. The former (the head region) has been cut off by the longitudinal valley and by its continuation in the south — the interior basin of Tuineje — Gran Tarajal, whereas the mouth has disappeared owing to the formerly severe attack by the ocean waves: an abrasion coast has been formed here (aided by faultings?).

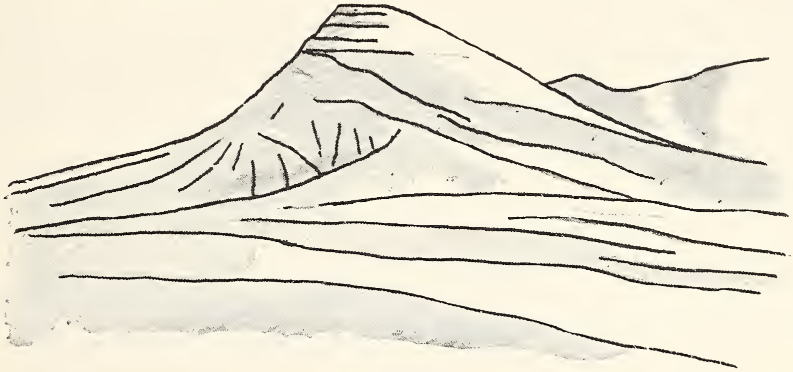

Fig. 7. Montaña de la Fortaleza (596 m) composed by the basalts of the table-land series with conglomerates at the base. Foreground: amygdaloid lavas. Looking north.

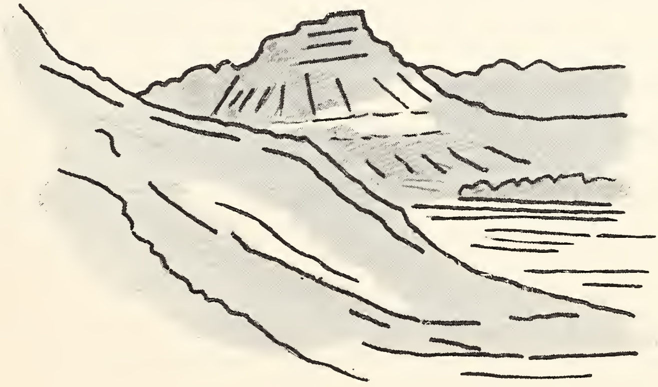

Fig. 8. Montaña de los Helechos (450 m) with Cerro de la Goma in the left foregr. Remnants of the basaltic table-land formation. To the right the end of a malpais-lava derived from Caldera de Gairia. Looking southeast.

All these cross-going valleys are consequently of rudimentary nature and represent so called »water gaps» (more or less perfectly formed). There are in the upper end of some of these valleys tresholds of insignificant height, but they do not disturb the general impression of a cross-going valley.

The mature stage of these valley rudiments points to a long period of erosion by rivers of much persistency. Certainly there lie off the eastern, southeastern and southern coasts huge masses of detracted fluviatile material, and the soundings in this area have also demonstrated the presence of a submarine platform.

In fig. 8, plate V, a panorama view is presented of one of the rudimentary valleys — Barranco de Aceituno — lying some km to the west from Gran Tarajal. This barranco (valley), which is directed to the south, has a low treshold, as is seen in the background of the picture. On the other side of the same there opens the wide basin of Tuineje, but the valley points to a former continuation across that basin towards a northern highland. We note the broad cross-profile of the valley. Of a similar aspect are many of the other cross-going valleys in the eastern Cordillera (fig. 9).

There are as far as the author knows in the Canaries no other remnants of an old drainage system of such an advanced maturity.

5. Faultings and erosion

After a period of long tectonic inactivity the former table-land now altered to a valley-dissected landscape with east- and southeast (and south) trending drainage pattern was affected by crustal movements along greater and smaller fault lines. It has not been possible to follow all these lines in the nature, along which the displacements have occurred. It is probable that some of the larger faults now are hidden beneath the sea. But there are in the present landscape of Fuerteventura some lines fairly well expressed. One such line of displacements can be followed from the vicinity of Oliva in the north to Gran Tarajal in the south for a distance of about 40 km. This line marks the western limit of Cordillera oriental. Another approximately parallel fault line runs somewhat more to the west and marks the eastern rim of the western Cordillera or Montañas de Betancuria. This line is much shorter than the former when following the longitudinal valley enclosed by them both. But the westerly line then turns suddenly to the west, describing a circular bow around an interior basin and finally runs east to the vicinity of Gran Tarajal.

These are the master faults that can be recognized. But we have certainly also many other faults of minor importance and of varying directions. They are, however, hardly to be fixed in the field. Perhaps the many transverse valleys dissecting the eastern Cordillera may have been controlled by such minor faults. There are likewise in the western Cordillera several smaller lines of fracture supposed to exist, but the construction of the fracture pattern seems to be a very difficult if not impossible task.

In short we have to imagine that the old eroded table-land was dissected in many directions, and a horst relief was created. Two great horst blocks were formed: the eastern and the western Cordillera. Between them there was opened the longitudinal valley, a very conspicuous feature of the island. This valley runs straight N—S bordered by steep mountain sides. Its bottom was filled with debris. Later on there appeared, as we will find, lava volcanoes in the middle sector, interrupting the continuity of the valley.

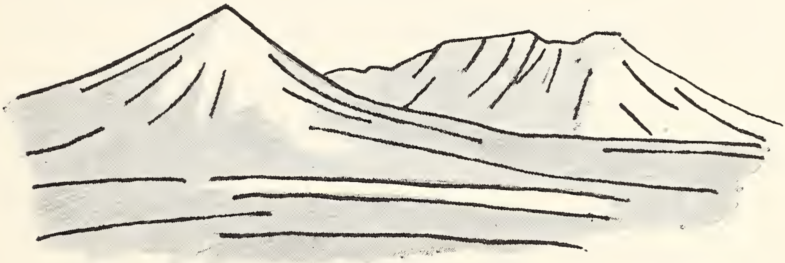

Fig. 9. Cerro Sombrerito (302 m) and to the right Montaña Aguda (494 m) remnants of the basaltic table-land series demonstrating a very advanced erosion stage in the eastern Cordillera. Looking southeast.

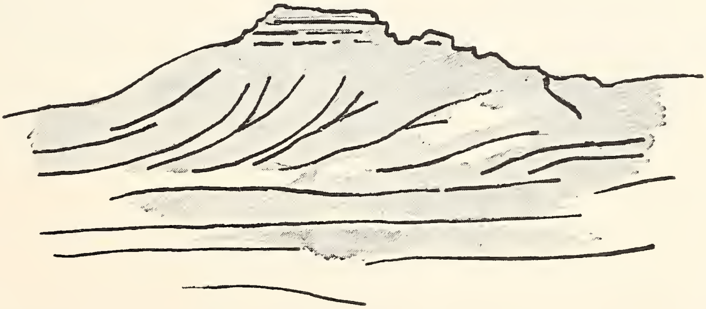

Fig. 10. Montaña de Vigán (487 m) in the southern part of Cordillera oriental, consists of the table-land series. Looking north.

The conditions ruling the peninsula of Jandía have not been studied by the author, but it is possible that a line of fracture runs along the north-western coast, likewise along the south-eastern coast (HARTUNG 1857).

The horst relief created by these events immediately provoked the revival of the erosion, and the conditions were rapidly changed. The western horst mountains — the original highland of the table land surface — finally lost their capping basaltic series, and the old basement was laid bare except where some remnants (witnesses) — Cerro Melindraga and Cuchillo Negro in the region of Chilegua — have persisted.

The eastern horst mountains (with their transverse valleys) consisting entirely of the table-land lavas, were of course also much eroded. The fact that a less resistant »breccia»-formation underlies the hard basalt banks was the cause of the formation of so many mesa — mountains (»*fortalezas*» as the islanders would call them) with nearly vertical precipices in the mighty lava banks (fig. 7, 8, 10 and 12).

The erosion in the western mountains progressed considerably also after the disappearance of the table-land series, and a drainage pattern was created of a kind entirely different from that of the table-land. A watershed came into existence close to the eastern edge of the mountains (at the longitudinal valley) and the barrancos were directed chiefly to the west coast. As time vent on these barrancos or gorges were widened to valleys, and the gradient of the bottoms was lowered. The intervening ridges were also lowered and rounded off. With the advanced maturituy of the relief the transporting power of the rivers diminished, and much material was left in the valley-bottoms.

The erosion on the inner (eastern) side of these mountains found a somewhat steeper gradient, hence the barrancos leading to the longitudinal valley got the character of gorges.

The northern part of the western horst mountains has forms of more advanced maturity than the bulk of the mountains.

With these changes of the surface conditions the general aspect of the island assumed features which still exist if we make an exception for the new constructive forms to be described in the following: the shield-volcanoes in the lowlands and the cinder cones strewn over the island.

6. Lava volcanoes in the valley bottoms

A special geologic feature in Fuerteventura not noticed by earlier travellers are the shield-volcanoes of the »Hawaiian type» met with in some parts of the interior lowlands. They have sent large streams of very fluid lavas of basaltic composition into different directions filling the bottoms of the valleys and basins repeatedly. Some of the lava currents have reached the coast in the west and in the east.

I have found altogether five (or six) such shield-volcanoes lying in the central parts of the island. Two of these are in the vicinity of Antigua—Triquivijate, one is to be seen in the immediate surroundings of the little village Casillas del Angel, at the head of the broad valley used by Rio de Las Cabras. Still another lava volcano lies on the way from Puerto de Cabras to Tetir, where the new airport has been opened. And at last we will find a smaller

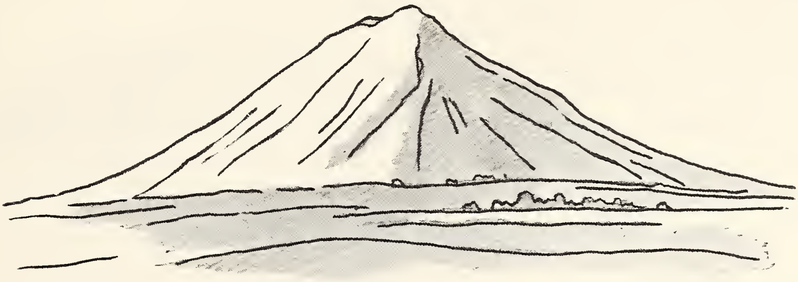

Fig. 11. Montaña de Tindaya (397 m) a conical trachyte mountain representing a volcanic plug laid free by the denudation of the brittle rocks of the table-land series once enclosing the plug. Looking west.

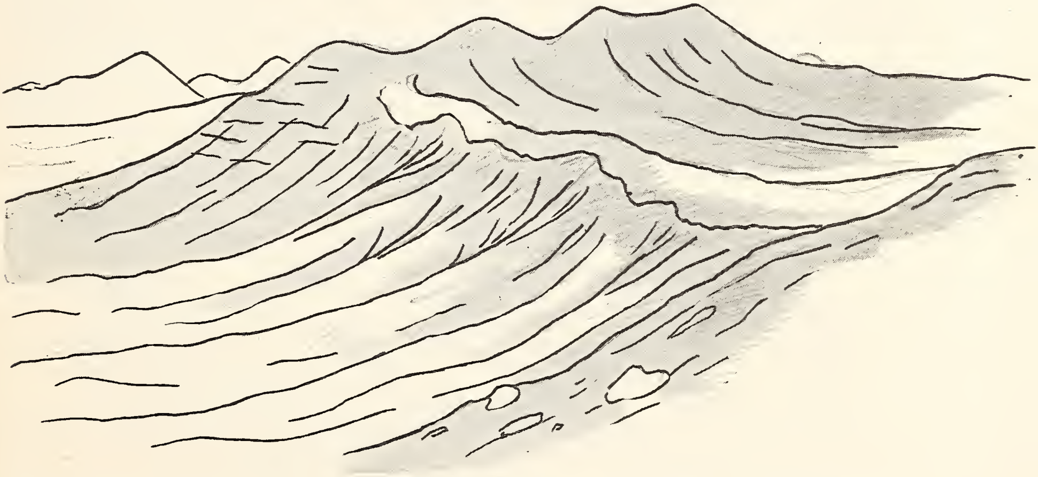

Fig. 12. The western edge of the basaltic table-land series, represented by a fault scarp at the eastern side of the longitudinal valley. Basin of Tetir to the right. A young cinder cone in the distance to the left. Looking north.

shield-volcano also on the opposite side of the island S from Mont. de Tindaya, to the NW from Montaña del Tao.

The orifices of emission of the copious quantities of lavas can not be more precisely fixed. The volcanoes are weathered and moreover covered with the »armour plate» of travertine, and finally also by the lateritic soil.

The greatest centers of lava emissions were those of Antigua — Triquivijate (fig. 15). Here lava accumulation was so great that a treshold in the longitudinal valley was formed, interrupting its continuity. Large floods of lavas ran down into opposite directions: to the north and to the south. The southern lava streams turned southeast and east-south-east and filled the transverse Valle de la Torre, ending at the sea shore. The northern branch covered a large lowland surface and then escaped to the west coast by way of Valle de los Molinos.

The shield-volcano at Casillas del Angel (on the way from Ampuyenta to Puerto de Cabras) emitted similar lava floods down the old transverse valley now used by the (intermittent) Rio de Las Cabras (fig. 16).

Of other shield-volcanoes we have one of larger dimensions lying as was said on the way from Puerto Cabras to Tetir, Llano del Viso. The flatness of this volcano makes it scarcely perceptible in the field; it has, however, sent copious masses of lavas down the coastal declivities in this sector of the east side of the island.

Another (small) shield-volcano is met with close to the northern highway, La Ventosilla, to the south from the trachyte mountain Tindaya. It has sent lavas down Barranco de Jarubio to the west coast. Finally there is a (dubious) shield-volcano in the vicinity og Montaña Bermeja, Montañeta de Tao. Lavas emitted from here may have been united with those arriving from the great center Antigua — Triquivijate.

As we have found this volcanic phase was of considerable importance in modifying the fault- and erosion relief previously existing.

7. The island emerges from the Mid-Tertiary sea. Limestone deposits

After the cessation of the lava outpourings from the shield-volcanoes just enumerated there elapsed certainly a considerable time span, during which the denudation forces worked along the coasts and in the interior of the island. The land surface was apparently subjected to a subsidence into the Mid-Tertiary sea, like several of the other Canaries, as has been proved.

During the inundation by the sea calcareous matter was deposited on the sea-floor in the form of foraminiferan ooze, accumulating to bottom layers of considerable thickness. The ooze was pressed together to a rather coherent lime sediment, a shelf deposit in water of probably some hundred meters in depth.

But as the time vent on there began a slow emergence of the island, and the lime sediments of the shelf were raised above the sea level. It has remained in that position up to the Recent time, forming a fringe along the coasts, later on however, much interrupted by erosion. The height of the lime sediments is

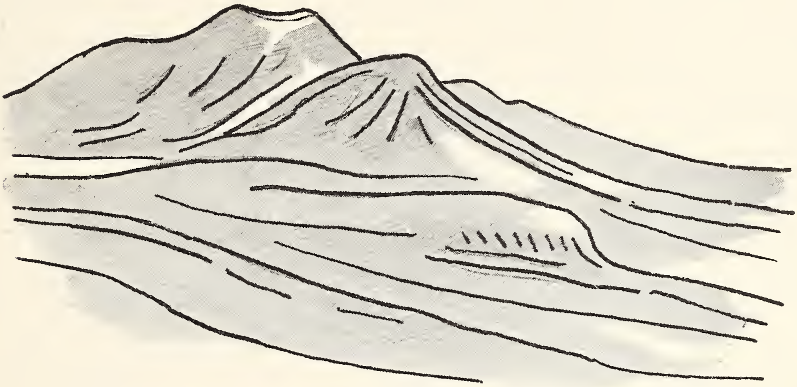

Fig. 13. The eastern fault scarp of the longitudinal valley looking south with a cinder cone grown up against the old basalts of Montaña Aceitunal. The cone is half destroyed by erosion.

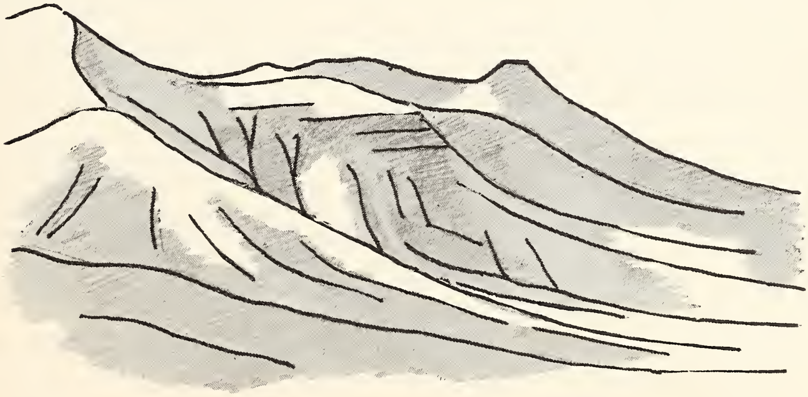

Fig. 14. The western declivities of Montaña Aceitunal, showing large exposures of the eastward tilted basaltic table land series. Looking south.

somewhat varying, but never more than ten meters. — The original ooze has been transformed into a hard yellowish-white stratified limestone of remarkable purity. The rock is now quarried for burned lime, which is exported to the neighbouring islands together with the copious masses of travertine limestone (burnt or not burnt).

A very appropriate locality for the study of the limestone in question is at Playa de Ugán, not far below Cortijo de Chilegua on the *barlovento* coast (see the fig. 10, plate V). The limestone rests here as is seen from the picture, in horizontal position.

The limestone is covering a very low platform of the basement rocks, and the total thickness measures about 10 m. This may, however, not be the original one, since it can be assumed that the erosion has carried away some of the upper layers (the mass is stratified). The limestone does not rest immediately on the old rocks, but there is at the base a conglomerate layer grading upward into a gray limestone (the lava grains are here mingled with minute particles of lime, fragments of foraminiferas etc.).

There are several such limestone capped terrace remnants along the *barlovento* coast, but the author has not visited them. The long distance from Puerto de Cabras has hitherto prevented the quarrying of these occurrences.

Since this limestone rests in parts on the table-land basalts, it is of later date than this volcanic cycle. To fix its age more closely is, however, difficult, if we intend to fit its formation into the special time scale of Fuerteventura.

On the east coast of the island, close to the mouths of Barranco de Herradura (north from Puerto de Cabras) and of Barranco de Rio de las Cabras (south from the Capital) one can see, how basaltic lava banks overlie a bottom stratum consisting of a yellowish-white limestone of not nearer known thickness. The layer lies only some m above the sea level. If this limestone corresponds to that of Playa de Ugan, its deposition in the form of clacareous ooze must have taken place at least before the flow of the basalt lava streams, which overran the limestone. The lavas in which Barranco de Herradura has been eroded belong to the shield-volcano Llano de Viso, and the lavas at the mouth of Rio las Cabras derived from the shield-volcano Casillas del Angel. Hence it seems that the eastern lime occurrences are of an age antedating the shield-volcanoes. It is, however, not proved, if the eastern limestone occurrences really are contemporaneous with the western ones.

In this connection it may be of interest to describe another limestone occurrence, this time on the *barlovento* coast in the vicinity of Playa de los Molinos. The locality is really very instructive in stratigraphic sense. — The lower course of Barranco de los Molinos is a deep canyon eroded in a series of horizontally lying basalt lava beds (9 beds) ending at the sea with a vertical precipice of about 30 m (fig. 17). This series of lavas has been emitted from the great shield-volcano N from Antigua, and they underlie also the wide basin Llano de la Concepcion. The lavas rest on a pale-yellowish limestone (or calcareous sandstone) which latter is exposed in a small barranco to the south from Playa de los Molinos. The approximative thickness here is about 15 m. The limestone lies on an old conglomerate which in its turn rests on the upturned edges of the Trapp-formation. The limestone appears also in the walls of the canyon of los Molinos forming a basement to the overlying thick basaltic lava beds. It is, however, apparent that the limestone along this barranco has been much eroded before the lavas ran down to the sea. The lavas have used the course of an old valley eroded in the limestone formation.

According to these observations we have to suppose that there existed in this sector of the *barlovento* coast a limestone formation of considerable extension also seawards, but later on attacked by the sea and by the fluviatile erosion. A valley was formed. In a later time the repeated effusions of basaltic lavas occurred, belonging to the shield-volcanoes in the interior.

Since no macrofossils have been found in the limestone deposits and the rock itself hitherto not has been examined, it is impossible to say anything

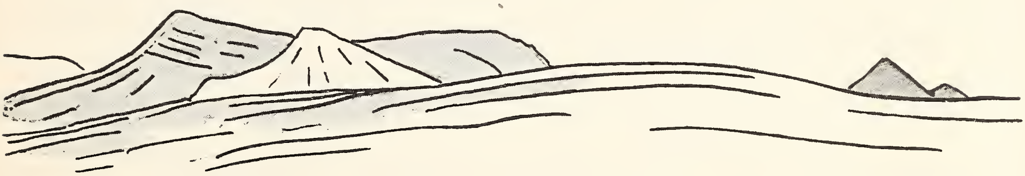

Fig. 15. The great shield- (lava-) volcano rising between Antigua and Triquivijate, in the middle sector of the longitudinal valley. Cerro del Campo (543 m) to the extreme left, forelaid by a half-destroyed cinder cone. Looking east.

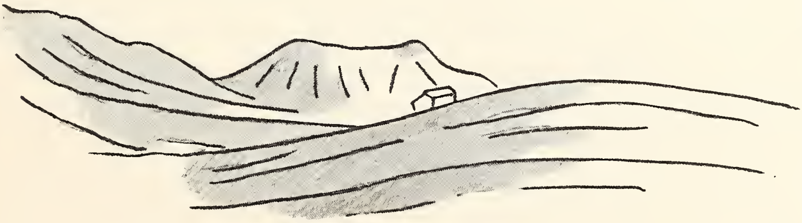

Fig. 16. The shield-volcano of Casillas del Angel, head region of Valle del Rio de las Cabras. The mountain in the background belongs to the table-land series. Looking east.

about its geologic age and its relation to the limestone of Playa Ugán. The only statement to be made is that the deposit is older than the repeated lava effusions from the shield-volcanoes.

The limestone is in its turn younger than the coarse conglomerate resting on the Trapp-formation, which is exposed in the vicinity.

Another limestone occurrence is met with at Puerto de la Peñita on the same coast somewhat more to the south. The author did not have occasion to visit this locality, which is mentioned by Bourcart (1938).

It is to be hoped that all these occurrences and their rocks may be more closely investigated in the future, since their stratigraphic conditions are of much importance. The same can be said about similar occurrences in the northern neighbour island of Lanzarote.

8. Late-Tertiary and Quaternary volcanic activity

In spite of the fact that Fuerteventura has in the main the stamp of an old erosion surface, in this sense contrasting against the other islands in the Canaries, there are nevertheless many volcanic cones with well conserved shape, several of them being of rather considerable dimensions. But comparing the frequency of such cones with that dominating the landscape in Lanzarote, they may be rather subordinate morphologic elements.

The surface of Fuerteventura is provided with at least 20 volcanic cones rather unevenly distributed. They are all of the so called Strombolian type: the cones consist of slags, lapillis and ashes, and they have mostly a wide crater, open in the northern sector. The surroundings of these cones are strewn with lapillis and ashes (mostly decomposed) and also bombs and stones. Lava streams have emanated from these cones, the younger of which now form the *malpaises*, inhospitable, rugged fields of glassy lavas (mostly *aa* — lavas). Some of these lava fields are rather extensive, as is the case in the extreme north and in the center of the island (here: Malpais Grande and Malpais Chico).

We can divide these cones into at least two groups according to their relative age. One category is certainly from Pre-Quaternary time, most probably from the post-Miocene period, the other cones are Quaternary in age, and some may perhaps have originated in Sub-Recent time. In every case there is no ground to assume that also volcanoes of the historic era are present, contrarily to the conception of von Buch. There are no traditions about such Recent outbursts, of which some should have been contemporaneous with the eruptions in Lanzarote in the XVIII:th century.

The Pre-Quaternary cones

We have no difficulty to separate these cones from the younger ones, because the former are generally somewhat eroded and attacked by the weathering agencies. The flanks of these cones are in parts capped by travertine, and the lime has also infiltrated into the scoriaceous and granular (lapilli) mass that composes the cones. Such phenomena can be studied here and there in the gravel pits or in gullies cut into the flanks. Also the lava streams belonging to this group of volcanoes have the stamp of relatively old age. The surface is weathered, and a bush vegetation has grown upon it. The corresponding ash fields in the surroundings of every of the cones have been tranformed into clayish (lateritic) soils and many times also eroded.

Good representatives of these relatively old cones are found in many parts of the island. If we begin in the north, we have on the east coast Montaña Roja

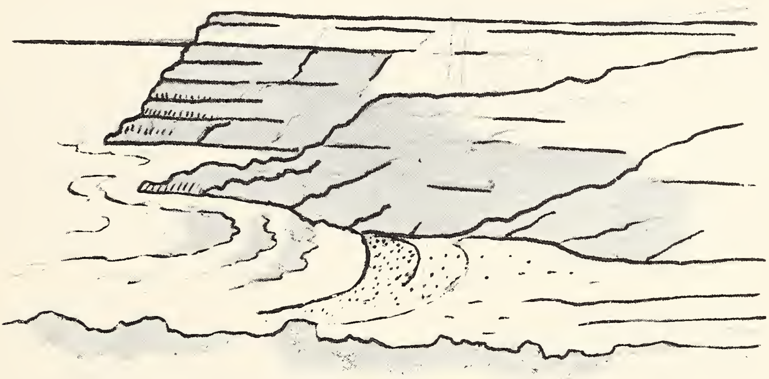

Fig. 17. Playa de Molinos, the mouth of Barr. de Molinos, *barlovento* coast. The abrasion cliffs expose a series of basaltic lava beds of formerly seawards wider extension. Looking north.

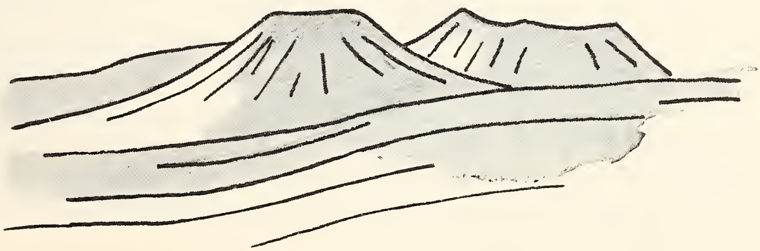

Fig. 18. Quaternary cinder cones, Cerro Negro (left) and Cerro Pajarito (right) in the northern part of the island, northeast from Villaverde. Looking northeast.

close to the sea, Montaña Quemada to the west from Matilla, Montaña Bermeja to the west from Pico de la Fortaleza, Montaña de Tamacite some km to the south from Tuineje, and Montaña Tirba, the two last named in the southern part of the island.

These old volcanoes have as was mentioned, ejected a fairly great quantity of fine particles of glassy material (ashes, lapilli) covering vast surfaces of the surrounding ground. These deposits now form the basis of the agricultural soils of Fuerteventura, at least the most fertile of these. Hence the economic importance of these outbursts cannot be underestimated.

The Quaternary (and Sub-Recent?) cones

We have now to consider a fairly great number of well conserved volcanoes, all belonging to a more recent period than the former ones, a fact which immediately can be recognized from the aspect of these volcanoes. They are all less weathered, the erosion in the flanks is still in the initial stage and the lavas are *malpaises.* A special feature of the cones is their wide caldera-like craters (see fig. 20, 21), mostly open in the nothern sector. Among these cones there are some of a rather fresh appearance, leaving one in doubt if they really are of Quaternary age; perhaps they must be placed into the Sub-Recent period (black cones and glassy *malpaises*).

Here is not the place to enter into a detailed description of the cones in question. We may shortly enumerate the most important of them.

If we again begin in the north, we have here a great group of rather imposing cones surrounded by vast *malpaises.* To the same assemblage belongs also the volcano on Isla de Lobos off the northeastern corner of the mainland. — The author has visited most of these volcanoes and also their caldera-like craters, all of semi-circular form. One of the most instructive of the volcanoes in this group is Montaña San Rafaél, the northernmost one with a wide caldera open to the north. Its bottom is occupied by a number of chilled lava bubbles. This volcano has emitted great masses of lavas into the northern direction, reaching in some parts also the northern sea shore. Of interest is a giant lava bubble, whose roof has collapsed. There are several other cones with semicircular calderas in the vicinity. The southernmorst of this group is Calderón Hondo, from where copious masses of lavas have expanded to the coast covering the northwestern corner of the island. Also ashes were blown out but in opposite directions.

A remarkable cone is Montaña de Arena some km to the north from Oliva (fig. 22). Its cone cosists of black coke-like lapillis (*picon*), and this is exploited in a big gravel pit in the southern flank. The lava is according to E. Jèrèmine (1938) an alkali basalt (basanitoid) and it has expanded over a wide area to the west and into the northern directions. A deep explosion crater lies in the northern side of the cone. Southwards the lavas have stopped just at the fringe of the Oliva plain (a clay filled basin). The eruptions seem to have occurred comparatively late, judging from the freshness of the lavas.

Of other volcanoes from the Quaternary period we find some farther south. The most remarkable (fig. 19) is Caldera de Gairia in the central part of the

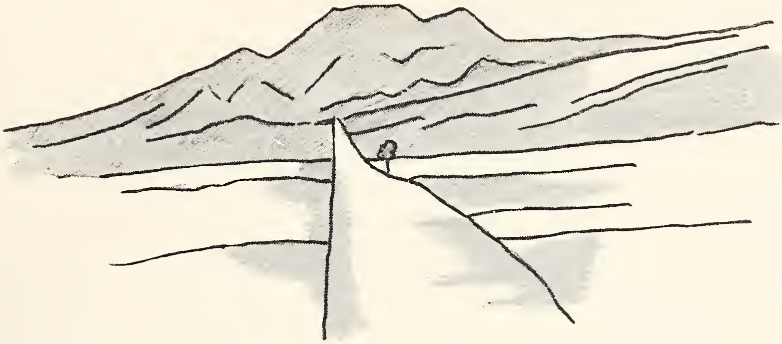

Fig. 19. Caldera de Gairia (480 m), seen from the north (highway Antigua — Tuineje). The cone consists of scoriae and slags. Lavas were emitted to the left. The cone is of Quaternary age.

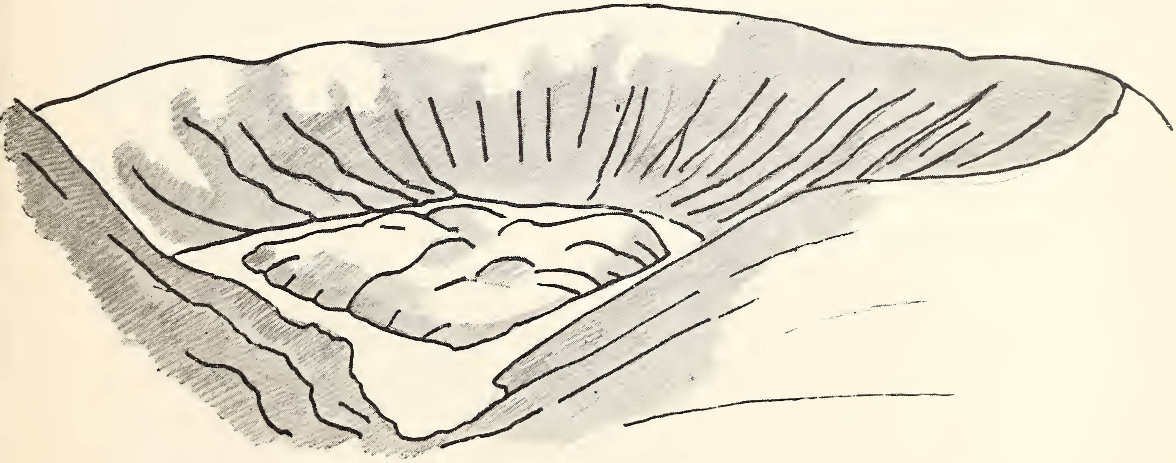

Fig. 20. Caldera de San Rafaél, the northernmost of the volcanoes of Quaternary age in the northern part of the island. Chilled lava bubbles in the bottom. Looking south.

island (to the south from Antigua). The author climbed the top of the caldera encircling wall (480 m), from which standpoint one gets an excellent panorama view over the surroundings. (cfr. with the sketch made by G. HARTUNG from the same place). The deep caldera-like crater (an explosion hollow) is open to the east, and through this gate there have escaped huge masses of lavas spreading over the flat country in the front of the opening. The lava is an olivine basalt, mostly glassy. The eruptions must have lasted for a long time judging from the massive cone built upon the old basement exposed here (in the bottom of the longitudinal valley).

Except Caldera de Gairia which is one of the most remarkable of the Fuerteventura volcanoes, there are in the vicinity three cones of rather fresh appearance: Caldera La Laguna, Caldera Liria and Caldera de los Arrabales. The two first named have sent copious masses of lavas to the east, forming a torrent which ran down the valley of Pozo Negro and stopped some hundred meters before reaching the east coast. This lava field is called Malpais Grande. The southernmost of the cones, Mont. de los Arrabales emitted a long arm of lava to the south, which stopped some km from Gran Tarajal.

The lava in question is an olivine basalt.

Maybe the three volcanoes enumerated are from the end of the Quaternary period or the Sub-Recent time. The freshness of their material is very apparent, and they seem to be younger than Caldera de Gairia.

Of relatively young volcanoes we have to mention a small cinder cone lying to the south from the Pájara village in the midst of the mountain region here. This cone is of a freshness similar to that of the cones just enumerated, and it has sent a lava stream down the creek that ends at Pájara, which lies in the bottom of a broad valley. The lava turned here NW following the course of the Pájara valley for some km until it stopped. This lava tongue rests on a fluviatile gravel bed that fills the valley bottom. The lava has a slaggy surface, but the interior is a compact grayish black lava (olivine basalt). The eruption occurred perhaps in Sub-Recent time.

No other cones are to be seen in the mountains of Betancuria.

The material collected from all the volcanoes enumerated in the foregoing is still not nearer investigated, hence no statements about their petrography can be made. The distribution of the cones may depend on the fracture lines dissecting the island. This is especially evident in the case when the cones are lying in the longitudinal valley, which is as we have found, a tectonic graben between horsts. The faults may have been reopened several times, so that lava could escape. Curious is the solitary appearance of the small Pájara volcano in a region that shows no other signs of more modern volcanism.

One gets the impression that the younger group of the volcanoes in Fuerteventura is more or less contemporaneous with the latest outburts in Gran Canaria, especially with the volcanoes forming the northern peninsula, which is called La Isleta. There are no traditions about the eruptions occurring here. The external aspect is, however, rather fresh both of the cones (not all) and their lavas and *picon.*

Some cones in the vicinity of Jinamar (on the way from Las Palmas to Telde); Gran Canaria, are no doubt of the same age-category.

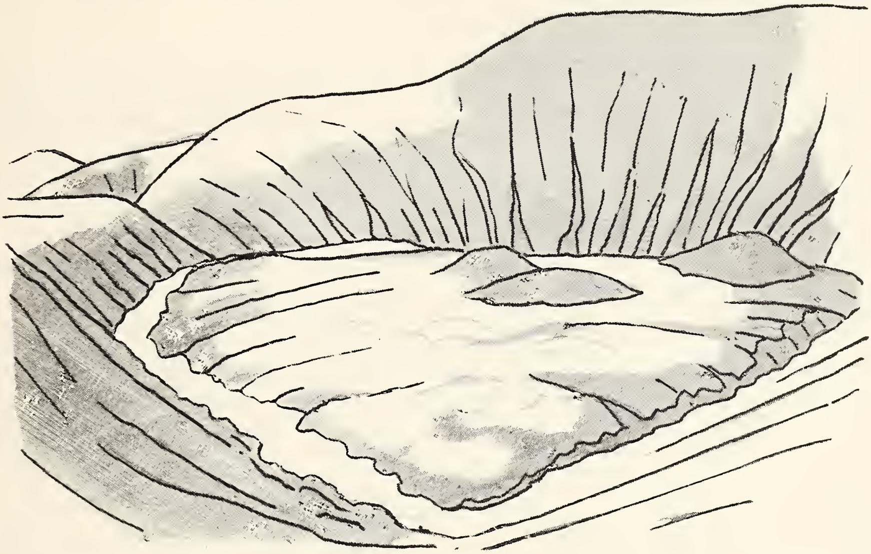

Fig. 21. A view from the northern rim into the caldera of the volcano Las Calderas in the northern part of the island. The volcano may be of Quaternary age and shows in the bottom chilled bubbles of lavas.

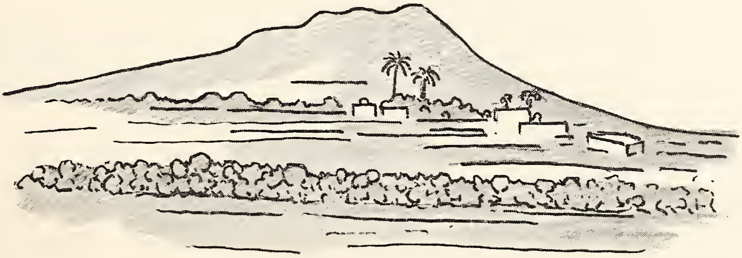

Fig. 22. Montaña de Arena (420 m), a Late-Quaternary volcano, region of Oliva. Looking north. Large lava streams in the surroundings have covered the brittle calcareous sandstone.

We have scarcely any reason to expect a revival of the volcanic activity in Fuerteventura, and the same can as it seems, be predicted concerning the island of Gran Canaria.

9. Littoral conglomerates

We may in this connection with some words characterize a number of occurrences of coarse conglomerates and gray sandstone layers met with along the east coast as well as in one place at the south coast (W from Gran Tarajal). They are at least in part of marine origin and seem to antedate the deposition of the travertine limestone.

If we follow the coast line from a point some km to the north from Puerto de Cabras to about 4 km to the south from the same place, we will find — especially at low tide — broad sheets of a coarse basaltic conglomerate with the boulders cemented by a sandy limestone matrix or by a calcareous dark sandstone. These sheets of conglomerate lie nearly horizontal or with a gentle dip toward the ocean. Sandstone underlies. Thickness is ab. $^1/_2$ m, some times more or less. Great pieces of the sheets have by the action of the surf been loosened and carried away, so that the underlying basalt is visible.

In the matrix there is sometimes an abundance of marine shells, mostly of *Patella* which, as is known, is common also among the shells of the present littoral fauna. In a locality S from the lime burning kilns S from Puerto de Cabras the abundance of the shells in the matrix is very apparent. The conglomerate is so firmly cemented that it can be blown into pieces only with the aid of a heavy hammer or by blasting. $4^1/_2$ km S from Puerto de Cabras in the direction of Matorral there is an occurrence of similar basal conglomerate grading into a gray sandstone layer which covers it. The layers are rich in shells of *Patella, Turritella,* also *Litothamnium* etc. The occurrence forms a plate in a nearly horizontal position and lies scarcely one meter above the tide water limit (Plate VII, fig. 13).

Going further in the southern direction and after passing Matorral (9 km fr. Puerto de Cabras) there are probably many such remnants of the conglomerate along the coast, which is a platform.

Between the mouth of Barranco de La Torre and Pozo Negro the path runs over several ridges and plateaus consisting of the table-land basaltic series but capped with a coarse conglomerate of up to 10 m thickness. The conglomerate has badly rounded boulders of basalt and is much impregnated with travertine lime. We have here formerly much more extensive gravel sediments. It seems to be a terrestric deposit which at the coast was abraded by the sea.

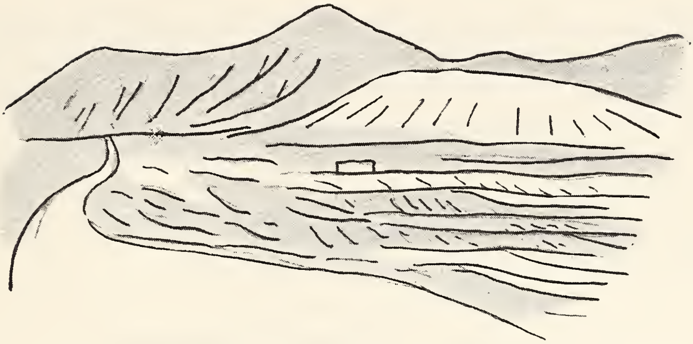

Fig. 23. Montaña de la Muda (689 m) highest point in the northern half of the island. Basaltic table-land series. Valle de Matilla in the foreground with pyroclastic deposits dissected by gullies. Looking north.

In the lower part of Valle del Aceituno to the W from Gran Tarajal there appears a very coarse conglomerate firmly diagenetized. It is a locality called Pasos de La Lapa. The conglomerate having the character of a fanglomerate rests on old basalts, and the boulders are also of basalts. The surface of the conglomerate deposit grades to the sea, but it is at the shore limited by an abrasion cliff several m in height. The conglomerate is capped with the travertine cover. Hence it is from pre — travertine time and probably of Late-Tertiary age.

As we have found the conglomerates enumerated are in part of littoral character.[1] In spite of their firm consistency they must be regarded as Late-Quaternary (or Sub-Recent?) in age, since the fossils are about the same as the living species. As we will see form the profile at Playa Blanca the calcareous sandstone is younger than this conglomerate, which marks a phase in the changes of the ocean level, when the island was slightly submerged. Then there was an upward movement, and the fine calcareous sand material was deposited as a terrestric product.

How far inland the sea shore migrated during the submergence has not been proved. Likewise it has not been established to what amount the island was raised in the subsequent time. In every case we will find that the present position of the litoral belt (between the high and the low tide limits) approximately coincides with that of the Late-Quaternary shore-zone.

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[1] On the south coast of the Jandia peninsula, at Morro Jable (finca of Gustaf Winter) there exists a layer of conglomerate or sandstone (?) containing fossils with *Strombus coronatus*, according to a kind communication of S. Benitez Padilla (Las Palmas). The locality is not nearer examined.

The inland conglomerate deposits in the form of scattered remnants are as we have seen, fanglomerates from a more remote period of mountain erosion, perhaps from the time following the horst movements discussed in the foregoing. It is only to wonder how little has in the reality been left of all such deposits of torrential kind. Much material may have been overrun by the basaltic lavas emitted from the volcanoes in different periods. Much has also been destroyed and furnished material to the Quaternary and Recent gravel masses in the valleys and the lowlands.

10. The travertine deposits on the land surface

A very typical formation of Fuerteventura is the nearly continuous superficial cover of a dense, concretionary limestone or travertine of varying thicknesses, sometimes attaining up to 20 m. The limestone, dense to the unaided eye, of a grayish-white to reddish colour is very hard with conchoidal fracture surfaces. It has in hand specimen a close resemblance to the known litographic stone of Southern Germany. The limestone of Fuerteventura is, however, never pure through the whole cover: only the superficial parts are of that kind, whereas in the downward direction stones are intermingled (embedded) in increasing number. Finally the limestone is to be characterized as a »breccia» with a cementing matter of lime and at the bottom it changes into the limefree bedrock. In vertical cuts the limestone shows a kind of banding indicating successive stages of formation of the superficial lime (fig. 13, plate VII).

The travertine cover occurs as was mentioned, nearly everywhere on the island surface, except where more recent erosion has destroyed the same. Further there are parts both in the interior of the island and along the coasts, where more recent gravels and sand deposits have capped the travertine.

Stratigraphically the travertine »armour plate» may be a good »leading horizon»: one can see which formations are older and which are younger than the travertine. Likewise one will be able to estimate the achievements of erosion in post-travertine time.

It is to be observed that the travertine is by no means a Recent product. Its formation has occurred under climatic conditions different from those prevailing in our days.

There has in the course of time been much discussion about the mode of formation of this travertine cover. The author has not the intention to quote here all the opinions expressed (starting with those of LEOPOLD VON BUCH 1825), but he will limit himself to his own conceptions on the matter.

The presence in Fuerteventura of such enormous masses of limestone: say 1.500 sq.km with an average thickness of (say only) one meter — corresponding to a quantity of ab. 3,8 milliards of tons seems quite unexplicable with regard to the composition of the underlying bedrock, which consists of basalts and their pyroclastic products — at least in the most cases.

We have in the foregoing described some occurrences of pure foraminiferan limestones along the coasts dating from times more remote than the period, to which the formation of travertine belongs. But even if we assume that this limestone formerly was more wide-spread than now, it is in every case strictly dependent on the vicinity to the coast, whereas the travertine is an universal cover and could scarcely have got its material from that source.

It remains therefore to seek the source of the lime in the bedrock itself, in the mineral components of the basalts. According to the analyses of basalts from Fuerteventura published in the paper of J. BOURCART and E. JÉRÉMINE (1938) the content of CaO varies from about 15 to 10 %. This lime content has apparently been dissolved by the circulating waters in the underground. A part of lime has been deposited as $CaCO_3$ in vesicles of the lavas (producing amygdaloidal rocks). Another part of the bicarbonate solution has been forced upwards along the innumerable cracks in the rock ground and finally reached the surface, where the carbon dioxide has escaped, leaving behind the insoluble $CaCO_3$ in the form of superficial incrustations. The efflorescence of the lime carbonate in this way postulates a more humid climate with markedly rainy seasons alternating with periods of intense sunshine. There must be supposed a rather strong evaporation in the surface to be able to raise the water solutions from the interior of the rock ground. That the efflorescence really occurred in a rythmic way, is seen from the fact that the travertine is built of a great number of concretionary sheets following the undulations of the ground. The alternance of wet and dry periods causing the rythmic deposition of the lime did not necessarily correspond to purely seasonal changes. One has also to think about long range changes of some irregularity.

It is evident that the formation of a lime cover that measures up to 10 or even 20 m thickness must have required a large time span several tens of thousands of years.

The travertine deposits have hitherto not been more closely investigated, and of chemical analyses the author has only one at hand quoted from the work of MANUEL CHAMORRO CUERVAS MONS (1952). Locality of the sample is not nearer indicated.

Travertine, Fuerteventura.

SiO_2	2.80 %
Al_2O_3	1.28 »
Fe_2O_3	0.84 »
CaO	48.38 »
MgO	3.20 »
SO_2	0.45 »
$CO_2 + H_2O$	43.05 »
$Na_2O + K_2O$	0.10 »
Sum:	100.10 %

Content of pure lime carbonate = 86.21 %

As we have pointed out the travertine limestone of Fuerteventura is exploited on an industrial scale. There are several quarries strewn over the nearer surroundings of the chief export base: Puerto de Cabras. The quarrying is going on mostly in an irrational way and with primitive tools. There are also several kilns, where the raw limestone is calcined, chiefly in the vicinity of the port. To a smaller amount also the foraminiferan limestone is exploited and transported by trucks from the distant occurrence at Playa de Ugán in the region of Chilegua.

11. Calcareous terrestric sandstone deposits

Now we have to turn our attention to another kind of sedimentary deposits in Fuerteventura, the role of which is not so important as that of the travertine. It is a terrestric evengrained and brittle sandstone with stratification sometimes scarcely perceptible; it is generally yellowish-white or rusty-yellow in colour and contains darker rusty streaks which are conform with the stratification. The evenness of the grains is sometimes disturbed by the intercalation of thin »conglomerate» layers, consisting entirely of thick, fingerformed petrified nests of the genus *Antophora,* rather firmly cemented together.

The sandstone in question consists of calcareous grains cemented by lime. Many of the grains are pulverized shells of marine organisms mingled with pulverized travertine. The rock should therefore be called a limestone, although the name 'calcareous sandstone' seems more appropriate.

The sediment has been deposited chiefly from the air, not from water. That is evident from the evenness of the grain, the size of the grain and from the whole appearance in the field. Of interest are the terrestric fossils enclosed in the rock. Except the enormous masses of the nests of *Antophora* we have shells of *Helix* and *Stenogyra* connected with fresh-water deposits. The masses of the *Antophora*-nests speak in favour of a more rich vegetation and consequently of a climate somewhat more humid.

The occurrences in the field are not very continuous. The chief region of sandstone deposits lies in the north. Here (to the N from Oliva) we have vast surfaces underlain by the calcareous sandstone covered with its own detrital product; the drifting calcareous sand. Owing to this sand cover the sandstone is seldom exposed to the daylight but forms the underground of the lowlands. It appears, however, in some places. We will find sandstone in Corralejos at the north coast. A sandstone cuesta is met with in the region of Los Lajares, at a locality called Los Jablitos (about 100 m above the sea). Here also a quarry is situated. The sandstone has in this cuesta a remarkable firm consistency compared with that of most other occurrences.

Good exposures of the sandstone in question are met with in some canyons eroded in the same. Around the gabbro-mountain Montaña de la Blanca (travertine covered) the depressions are filled with the sandstone to a thickness of at least 10 m and well exposed in the vertical walls of the canyons. Similar sandstone deposits are met with in the western environments of the volcano Montaña de Arena, the lavas of which have overrun the sandstone.

Turning to the east coast we have in the valley between Calderas Blancas and Montaña Roja (two volcanoes of the older group) a rather mighty deposit of the brittle calcareous sandstone. It covers a slope grading to the coast and is much dissected by gullies. It is stratified and carries abundantly of shells of *Helix* and *Stenogyra*. The position above sea level is rather considerable — 100—150 m.

Of sandstone occurrences situated more to the south we have to mention a coast cliff some km to the south from Puerto de Cabras at a locality called Playa Blanca, previously described by J. Bourcart (1938). The cliff exposing sandstone with a conglomerate at the base is about 10 m in height and demonstrates that the former extension of the deposit was much wider; most of it has been destroyed by the sea. A closer examination of the profile of the cliff reveals the following: at the base which consists of basalt and is invisible in this locality (but appears some hundred m to the north as a low coast platform) there lies a coarse conglomerate with pebbles of basalt. This grades upward into a gray sandstone of a firm coherence. It grades in its turn into a brownish-yellow sandstone with vertical partings. This rock gets more and more loose in the upward direction, until it is capped by a hard concretionary (tosca — like) limestone. The lower part of the sandstone encloses numerous vertical pipes of different length and width. The dark colour of the lowermost part of the sandstone depends on the admixture of minute grains of basalt. In the upper pale-coloured part of the sandstone bluff there is a distinct cross-bedding (petrified dune material).

The marine abrasion has as was mentioned been responsible for the reduction of the deposit to a small remnant and also for the copious masses of drifting sand in the southern direction from the occurrence; likewise for the broad playa of fine sand (Playa Blanca) existing here.

Of other occurrenes of such sediments the author supposes the existence of a yellowish sandstone at the *barlovento* coast close to Jable de Salinas (with »Risco Blanco»).

Moreover there must be similar deposits in Istmo de la Pared at Matas Blancas judging from the huge masses of drifting sand in this region.

The occurrences enumerated demonstrate that the deposits in question all are located in the low regions of the island and especially where they have been sheltered against later erosion. If lying on the coast the deposits have been subjected to marine destruction in a rapid tempo (Playa Blanca). The geologic age seems to be Quaternary corresponding perhaps to a period of emergence from the ocean and of a somewhat more humid climate. An emergence would perhaps identify the period with a time of glaciation in higher latitudes (eustatic lowering of the ocean level). The stratigraphic sucsession at Playa Blanca (east coast) supports the conception of negative migration of the shore line at the time of the deposition of the fine sediments, followed by a submergence of the island with attack by the surf on the sandstone deposit at Playa Blanca.

12. Late-Quaternary and Sub-Recent erosion and accumulation

We may now look at the achievements of the exogenic forces on the island in the period following the deposition of the travertine cover: the erosion in the mountains and in the valley bottoms and the accumulations of the detracted material. Most records are of the running water. The calcareous sandstone deposits which also belong to the Quaternary period above described, considered in the main as eolic sediments are to be kept apart from the deposits now to be discussed.

Wandering in the mountain valleys and barrancos one will soon find that there has in the bottoms occurred a considerable deposition of gravels and sands resting on the travertine. The thickness of these valley trains can be seen in cases when later erosion has cut a canyon into the valley fill. I have seen walls of nearly 10 m altitude exposing badly stratified gravels and here and there sand layers, all enclosing shells of *Helix* and *Stenogyra* etc. These accumulations must be of Late-Quaternary age.

The basins on the island have been filled with sediments, mostly of finer grain. In the northern part of the longitudinal valley there has been deposited much alluvial material. Here existed in Quaternary time a lake which was drained to the sea by way of Barranco de los Molinos. As this barranco was deepened by erosion, the lake was drained off, and a dry basin was left behind with its silt sediments containing also gypsum.

Other basins have been filled with clay sediments. That was the case for inst. with the Oliva basin in the north, where the little village Oliva is situated. Another basin is more kettle shaped — the basin (*cuenca*) of Tetir, also enclosing clay sediments. These clay products were washed down from the surrounding mountains (see fig. 12).

Gravel and sand accumulations have filled all the bottoms of the old valleys in the eastern Cordillera and in the basin of Tuineje — Gran Tarajal.

To the Late-Quaternary time seem to belong also all the canyons eroded in the lava filled valley bottoms (Barranco de Jarubio. Barranco de los Molinos, Barranco de la Torre), although the erosion has continued in the Sub-Recent time. Easily eroded were the deposits of the calcareous sandstone. Here narrow canyons several m deep have been opened together with complicated systems of gullies (bad lands in miniature). The thick deposits of volcanic ashes and lapillis have likewise been intensely dissected, as is the case in the bottom of Valle de Matilla (fig. 23). These deposits were before their erosion altered to a lateritic soil.

Owing to the dry climate in the Recent time and up to the present day no perennial rivers have existed here since the Quaternary period. It is therefore understandable that the surface of the island has not the stamp of countries with rains all the year. Weathering and wind transport have to a certain degree obliterated the erosion forms (which are mainly Quaternary in age).

Typical of the rivers here fed by the winter — cloudbursts are their copious masses of silts and clay washed down the slopes by the rain from all the weathering products covering the rock ground. The water of the swift, short-living currents is of a chocolate brown colour. These fine particles in suspension are sedimented farther down in the shallow basins or valley bottoms. But they do not come to rest for any long time.They are repeatedly translated toward the shore and finally catched by the breakers.

This transportation of silt and clay material has much hampered the use of the water storage dams constructed in some of the barrancos. The dam of Barranco de Herradura has been completely filled with these fine grained sediments dragged from the basin of Tetir. Also the new storage dam in Angostura La Peñita of Barranco de la Villa (below Vega Rio de las Palmas) is on the way to be filled with these sediments. In the largest of the dams situated in Barranco de los Molinos the filling process has not yet reached the critical stage but with time this dam cannot escape the same fate. Since the excavation of the silt and clay fill inside the dams may be a too expensive undertaking, the only solution seems to be an additive height of the dam. That is, however, in the long run a temporary measure.

In the island of Gran Canaria where the dam-storage procedure is extensively practized, no such temporary trasportation menacing the dams is to be feared. That depends on the fact that the land surface of Gran Canaria is too rapidly eroded to allow the formation of fine grained weathering products (laterite etc.).

Of accumulations from the Recent time there are to be mentioned also the drifting sands derived from the calcareous sandstone deposits. Such fields either barren or covered to some degree with shrubs have not very great distribution, and they will be shortly mentioned in the following.

13. Soils and ground water conditions

The soils of Fuerteventura have hitherto not been systematically investigated, and the only data being at the authors disposal are his own observations together with sparse notices found in the literature and informations obtained from persons resident on the island.

In the mountain areas we have the rock surfaces seldom exposed to a greater degree. In the crests and the *cumbres* there is mostly a rather thick cover of a dark-brownish, stone mingled soil. It is to wonder how this soft earth has been able to maintain its elevated position despite the winter rains which are relativelly most frequent in these altitudes. Lower down the slopes the travertine cover appears but it is generally covered with the stony earth, furrowed with gullies. Sometimes the travertine has been clean-washed and gives to the slopes a bright colour (»*Montanas Blancas*»).

In the lower regions, lower than the accumulation fringes of the piedmont gravels there are mostly fine-grained sediments when accumulated in basins, but gravels and coarse sands if sedimented in the trains of the barrancos. These soils are appropriate for the agriculture if ground water can be obtained by drillings (see in the following). Most important soils in the lowlands are, however, the lateritic clay fields covering the travertine to a thickness of $^1/_2$ m and less. This clay sediment is often brick red, sometimes dark brown and very hard in time of drought. Its chief occurrences are in the center of the island around the old Villa de Antigua, but also in the Tuineje region farther south. This soil is of high value to the farmers, but it must be blended with a more coarse grained material (*picon*) to get a higher degree of permeability.

The formation of this laterite demands its explication. As has been mentioned it is of post-travertine age. One has to think about fine volcanic ashes deposited over large surfaces. These sediments were partly destroyed by erosion, partly chemically altered under the climatic regimen of the Recent period (drought and winter rains alternating). The soil in question is consequently autochtonous.

Sometimes these originally pyroclastic sediments attain a considerable thickness, for instance in Valle de Matilla (N from Tetir), where it has been subjected to vigorous erosion and transformed to a »bad land» in miniature.

But there are also allochtonous clayish sediments in Fuerteventura although of small extension. The author has knowledge of two occurrences of clayish allochtonous basin fills: in the vicinity of Oliva and in the basin of Tetir.

The basin fill of Oliva has a surface plain as a table, and the village is built on it. This surface lies 250 m above the sea. The thickness has been proved by deep drilling for recovery of ground water and it attains one hundred m (?) Samples of this clay sediment had been sent to a chemical laboratory and analyzed (Woodcock & Mellers, London). The data obtained from one of the samples (depth not indicated) has been kindly communicated to the author by Mr Tomas Benjamin Dorta of Santa Cruz de Tenerife, a land owner in Fuerteventura:

Clay, Oliva

SiO_2	48.28 %
Al_2O_3	15.14 »
Fe_2O_3	10.20 »
CaO	6.72 »
MgO	2.58 »
SO_3	tr
Na_2O	1.76 »
K_2O	
CO_2	15.32 »
H_2O	
Sum:	100.00 %

Hence we find a very iron-rich product, and the original material was no doubt ashes from the surrounding basaltic volcanoes. These ashes were subjected to decomposition and then washed down to the basin from the slopes in the vicinity. The plain is not cultivated due to absence of water.

The thickness of the clay deposits of the basin of Tetir enclosed nearly on all sides by mountain slopes, seems to be considerable judging from the shafts dug for the search of ground water.

There is still another kind of sediments in some parts of the island surface: the drifting dune sands, derived from the soft calcareous sandstone deposits. These sand fields have been nearer described in an earlier chapter. They are of course very dry grounds. In some parts a bush vegetation — a *matorral* has grown up which is used as pastures for the sheeps and goats. — In later years the Spanish Ministery of Agriculture has here started (tentatively) the cultivation of the plant *Aloe perfoliata* to obtain the raw material for the fabrication of hempen. The plantations are run by using the dry farming method (the soil is covered with *picon*).

In general the soils of Fuerteventura are not bad for many kinds of agriculture. There are vast surfaces of a brownish soft soil, mingled with stones (mostly travertine fragments) covering the slopes grading towards the eastern coast which certainly could be cultivated if there were enough of irrigation waters. The sandy alluvial flats at the coasts, where shallow ground water can be obtained by way of wind motor pumps are to some parts laid under tomato plantations. Likewise one will find tomato plantations here and there on the vast surfaces of the flat basin in the southern part of the island, the basin of Tuineje — Gran Tarajàl. Here also the basal ground water is attainable by shafts and wind motor pumps. The soil is a mixture of sand and clay.

A rather serious question arises from the soil destruction by rain wash. Many tracts of fertile soils have already been badly damaged and transformed into badlands of narrow and deep gullies as was mentioned above (Valle de Matilla — Guisguey). The absence of a vegetative cover facilitates of course the destruction that progresses under every rain shower of some violence.

As far as the ground water in Fuerteventura is concerned the author has devoted some attention to its occurrence. We will find here quite different conditions from those prevailing in the islands of Gran Canaria and Tenerife — the chief agricultural regions. (H. Hausen 1954).

Since Fuerteventura suffers from a nearly continuous drought during most of the year, it is evident that the ground water resources here are far more restricted than in the mentioned islands. This statement does not, however, denie the possibility of some ground water accumulation in the basal parts of the island, especially in the vicinity of the piedmonts and along some of the valley bottoms.

The trade winds blowing from the northern sector during the mayor part of the year bring some moisture that is condensed from time to time, chiefly in the elevated parts of the mountains. The maximum of moisture is condensed in the winter months as rains, but also during the other seasons there is a direct condensation from the night clouds. The water condensed percolates into the stony soil or in the blocky rock surfaces of the mountains. The general absence of the rather impermeable travertine cover from the crests of the mountains permits a more easy infiltration into the ground. If we think of the western mountains (region of Betancuria with Cuesta de la Villa) the steeply dipping sheets of the Trapp formation facilitate the downward movement of the ground water, so that it finally attains rather low parts of the mountains, where it is accumulated as the basal ground water.

Montañas de Betancuria are in the reality the hydrographic center of Fuerteventura, where the condensation of the atmospheric moisture attains its maximum. From here a great number of valley courses originates, departing into different directions. Here also the ground water is accumulated in relatively mayor quantity forming the basal ground water center. No intervening impermeable layers or sheets of lavas impide the downward percolation eventually perching the water at higher levels.

The ground water thus accumulated moves slowly outwards from the mountains and seeks its way toward the coasts. Owing to the slow intake of atmospheric water the ground water currents in the basal water body are moving very slowly. In some sectors they perhaps never reach the sea shore. Many of the old valley bottoms have been utilized by the currents.

The water percolating through the mountain mass to its base is generally fresh. Also water recovered from the piedmont region by way of shafts (*pozos*) is of good quality or slightly saline. But when going farther away from the mountains one will find that the water of the (eventually present) shafts is getting more and more saline.

This depends on the special conditions prevailing in the lowlands. Here the ground is nearly entirely covered with the travertine limestone, which contains varying quantities of soluble salts (of calcium and magnesia etc.). When rain water percolates in such a ground it catches soluble salts, and the solutions are finally added to the basal ground water.

Favourable sites for irrigation with ground water are consequently the land fringes close to the mountains or in the valley trains inside the mountains. The villages Antigua, Santa Inez, Tuineje, Toto and Pájara are in favourable situations. Unfavourable are the regions of Oliva in the north (several shafts have proved dry) and the plains of Tefia.

It seems that the development of the ground water recovery in Fuerteventura is still in its initial stage. A more thoroughly going investigation is necessary, before we can state anything sure about the ground water levels and the movement of the water, its sal content etc. There have been dug, it is true, many shafts (*pozos*) in different parts of the island, but most of them have proved dry. It is of interest to note that the site of such operations was often determined with the aid of the divining rod (»*la varita*») in spite of the fact that the works in question are very expensive (diam. of the shaft 3—4 m, depts 50—100 m, shaft walls of solid stone masonry).

There exist now, it is true, a great number of shafts provided with wind motor pumps in different parts of the island, which produce ground water for the irrigation of alfalfa (luzern), tomatoes and maiz, but the water is generally not potable, except in some piedmont regions. Hence the culture of bananas is not possible under these circumstances.

The saline ground water recovered in this way seems to correspond not to the basal ground water body of the island, but to some secondary levels, where the water is perched by some impermeable tuff layers; Most of the shafts in operation do not reach to greater depths than 15—20 m, whereas the surface of such localities lies to one hundred and more meters above the sea. Hence it is clear that the main body of ground water must lie on deeper levels. Perhaps it is also less saline than what is the case on the secondary water levels. In the vicinity of the coasts the distance to the basal ground water level is insignificant and can easily be attained by drilling operations or by digging of shafts. Such is the case for instance in the region of Matorral (east coast).

One of the most promising regions of the island is according to the opinion of the author the southern basin Tuineje — Gran Tarajal. This tectonic depression, which is drained to the sea through the broad gate of Gran Tarajal, rises from here slowly to the periphery — toward the encircling mountains, to about $>$ 100 m altitude. The basin is provided with a very comprehensive ramification of barrancos of low gradient and with bottoms filled with gravel and sand. The low plateaus separating the barrancos one from another are covered with an arable soil of a brown stony earth. Shafts could here be dug in many places and water is not far below the bottoms of the barrancos. Nearer to the mountains the water is fresh (drink water is transported from a *pozo* here situated to a large farm in the vicinity of Gran Tarajal); getting more saline when going downhill, so that the *pozos* in the bottom of the outlet from the basin — the gate at Gran Tarajal — are all saline.

Wells and shafts in the interior parts of the basin in question attaining depths of ab. 100 m would certainly give sufficiently of water for the needs of new plantations of tomatoes or alfalfa (luzerne) in this region. Good starts have already been made by some land owners for instance by Mr TOMAS BENJAMIN DORTA in a lot called Diego Alonso.

Also in other parts of the island there should be made attempts to drill or to dig shafts to sufficiently deep levels, especially in the longitudinal valley and in the transverse valleys in the eastern Cordillera. In the lower course of the latter there would be possibilities to catch the basal ground water comparatively easily.

On the other hand in the north of the island the conditions seem to be less promising, as will be found from the numerous failures to get deep seated ground water by way of shafts.

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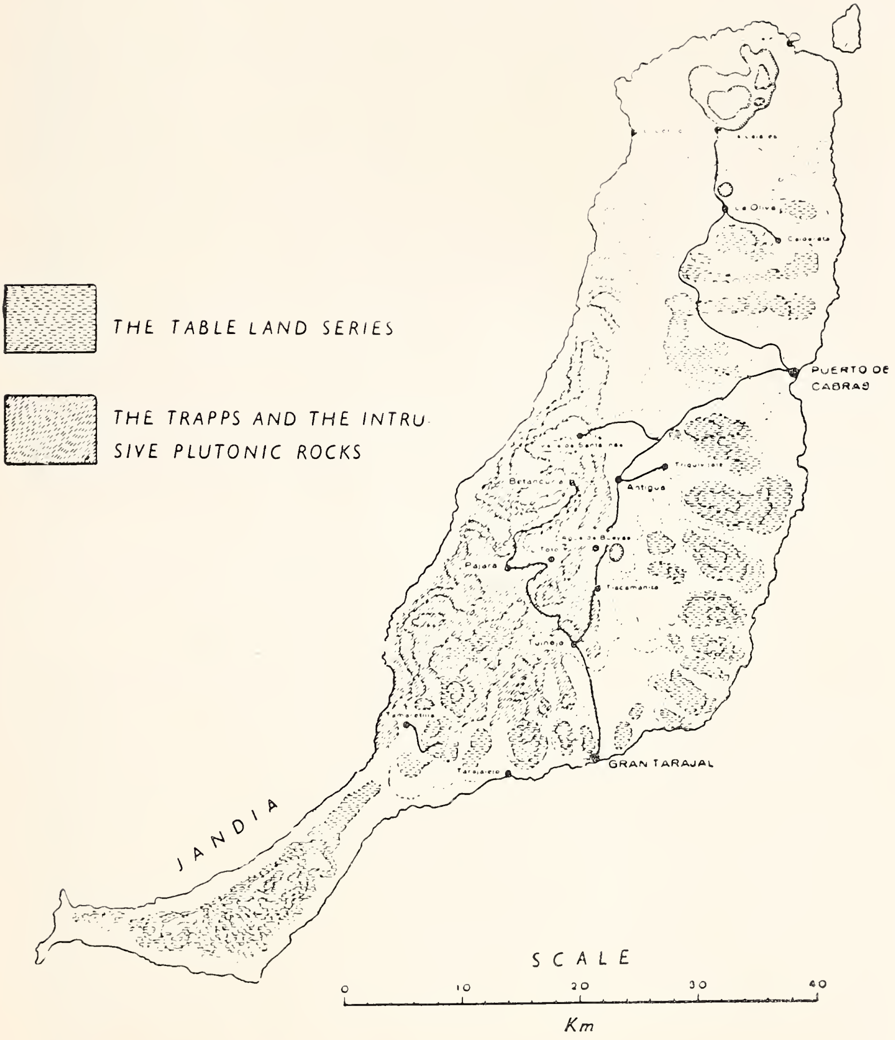

Fig. 1. Geologic sketch map of Fuerteventura. Areas left unsigned are lowlands covered with alluvium, the bedrock consisting mostly of basaltic lavas.

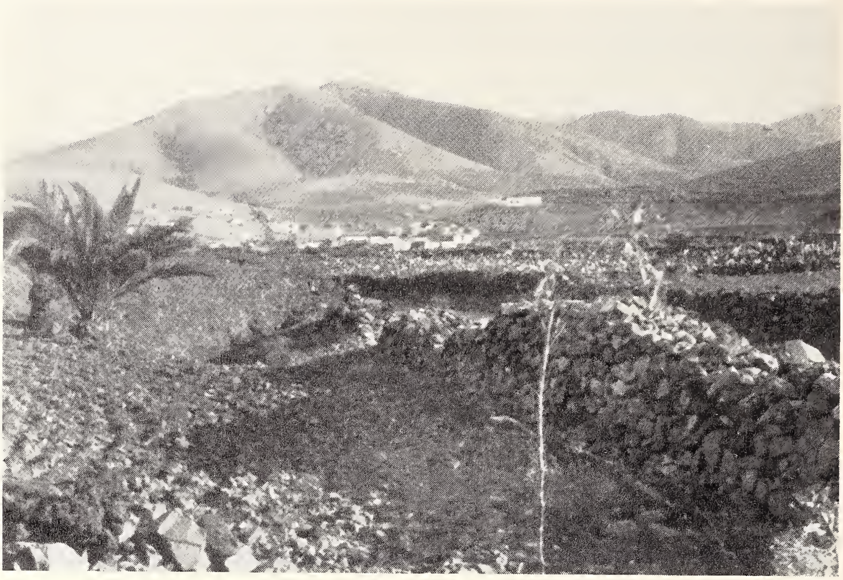

Fig. 2. Morro Fenduca (609 m) seen from the south. It is a part of the much eroded old mountain chain of the »Trapp-formation». Village Pajara at the foot of the mountain.

Photo H. Hausen 1954

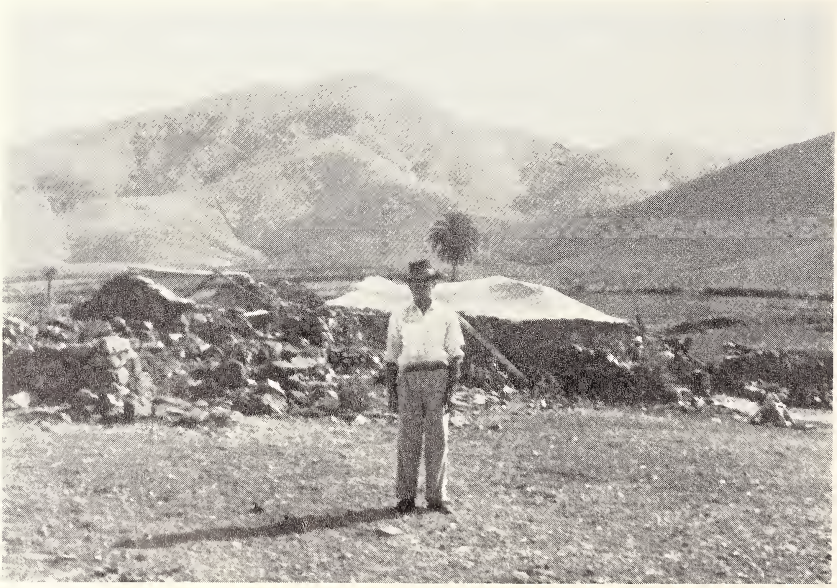

Fig. 3. Montaña Melindraga (619 m) a prominent peak in the region of Chilegua. The top consists of basalt resting on the upturned edges of the »Trapp-formation». Looking east from Fayagua.

Photo by H. Hausen 1953

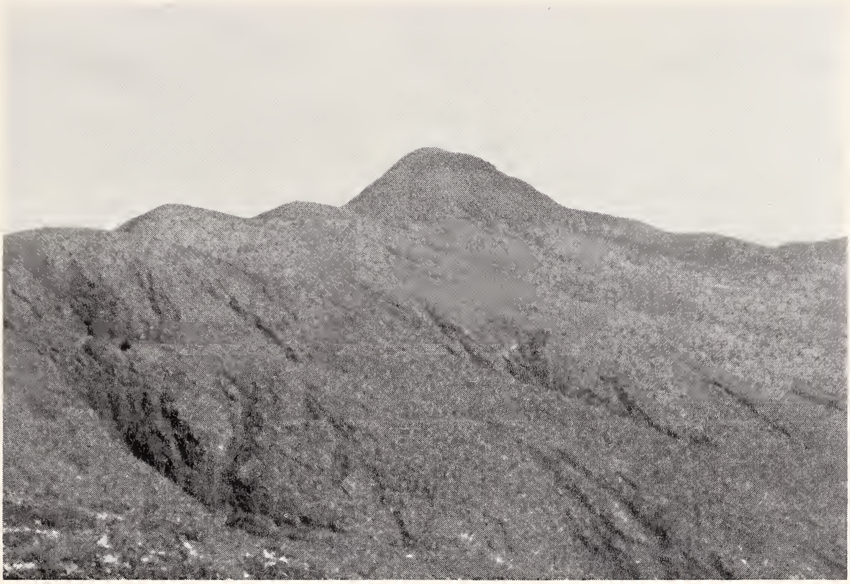

Fig. 4. Montaña Melindraga from the southwest (path from Chilegua to Fayagua). The cap of basalt — remnant of the old table-land formation — is well visible. Slopes in the foreground and middle distance: the folded "Trapp-formation".

Photo by H. Hausen 1953

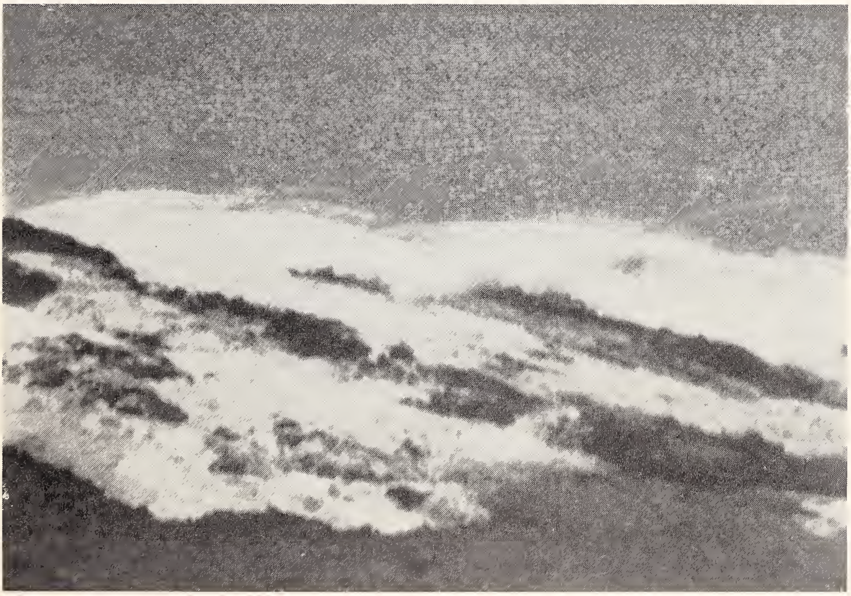

Fig. 5. The spurs of the folded Trapp-formation disappearing into the sea. *Barlovento* coast near Laja Blanca.

Photo by H. Hausen 1953

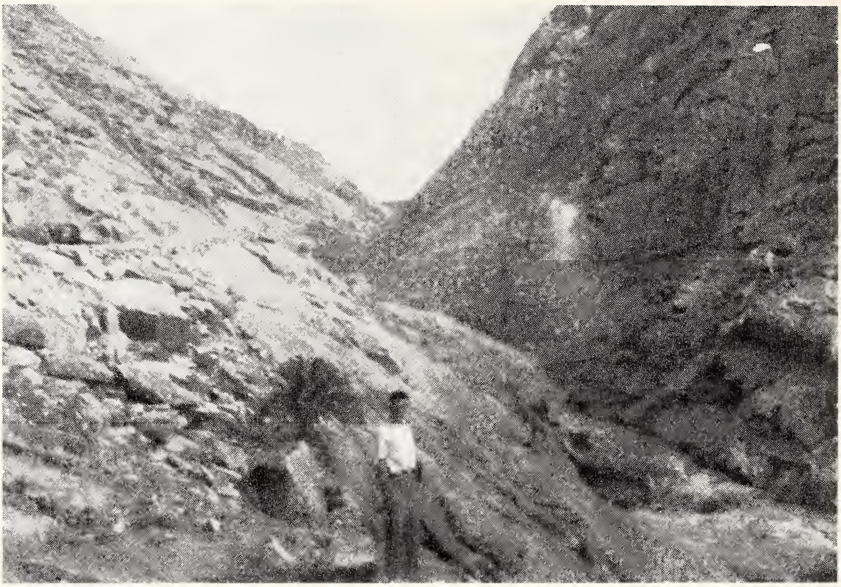

Fig. 6. The V-shaped erosion gap of La Peñita, Rio de las Palmas, a cut across the batholith of plutonic rocks in Montañas de Betancuria. Looking east (upstream).

Photo by H. Hausen 1953

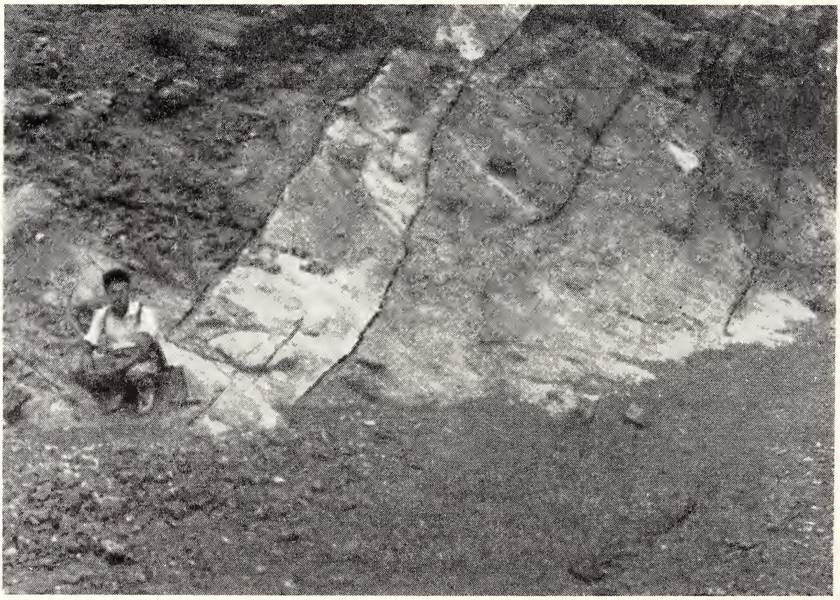

Fig. 7. A nearly concordant dike of trachyte in the upfolded Trapp-formation. Barranco La Peña, to the west from Betancuria. Looking north.

Photo by H. Hausen 1953

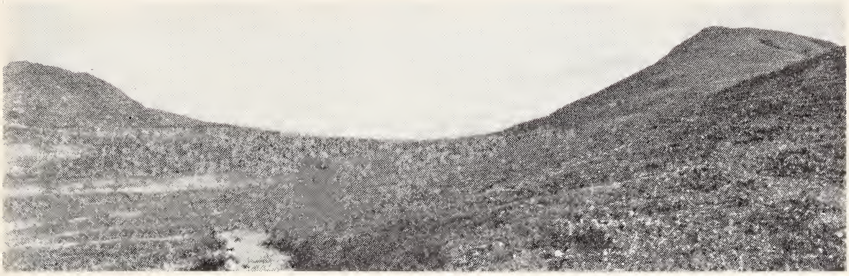

Fig. 8. A typical cross-profile of one of the mature valleys on the south coast with a treshold in the valley bottom. Barranco de Aceituno to the west from Gran Tarajal. Looking north.

Photo by H. Hausen 1954

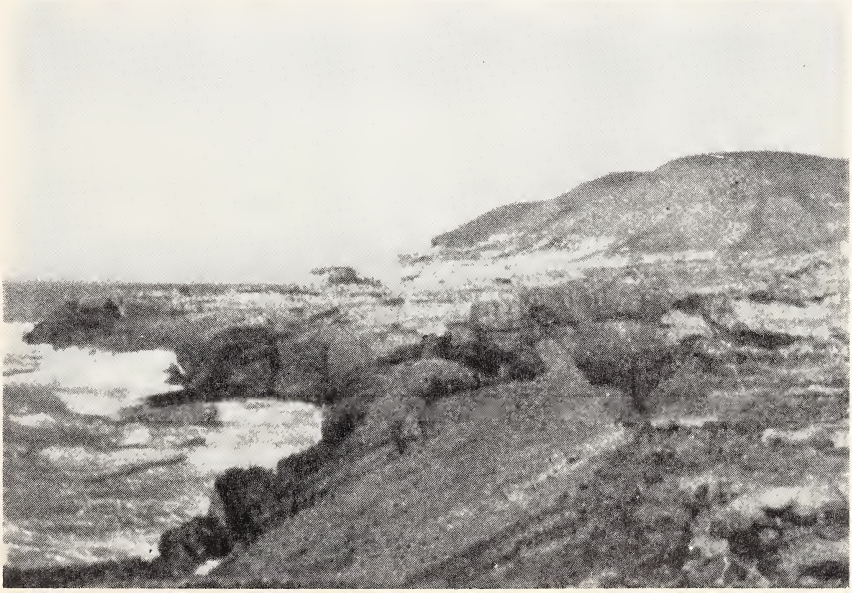

Fig. 9. A marine rock terrace on the *barlovento* coast, to the north from Puerto Nuevo, with a calcareous sandstone layer. The foraminiferan limestone which capped it has (except remnants) been worn away.

Photo by H. Hausen 1953

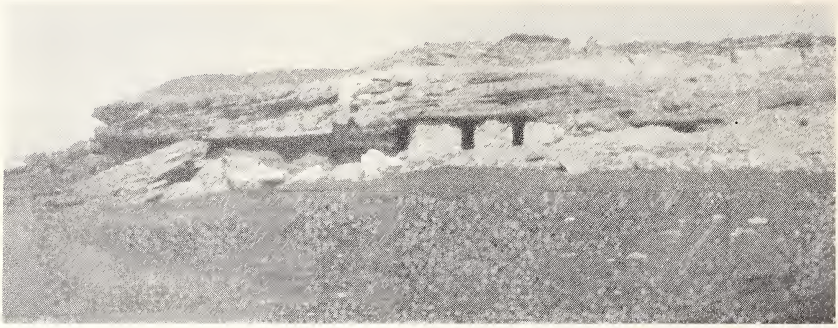

Fig. 10. A remnant of Tertiary foraminiferan limestone resting on a rock terrace some m above the high tide level. Quarry at Puerto Nuevo, Playa Ugán, *barlovento* coast. Looking north.

Photo by H. Hausen 1953

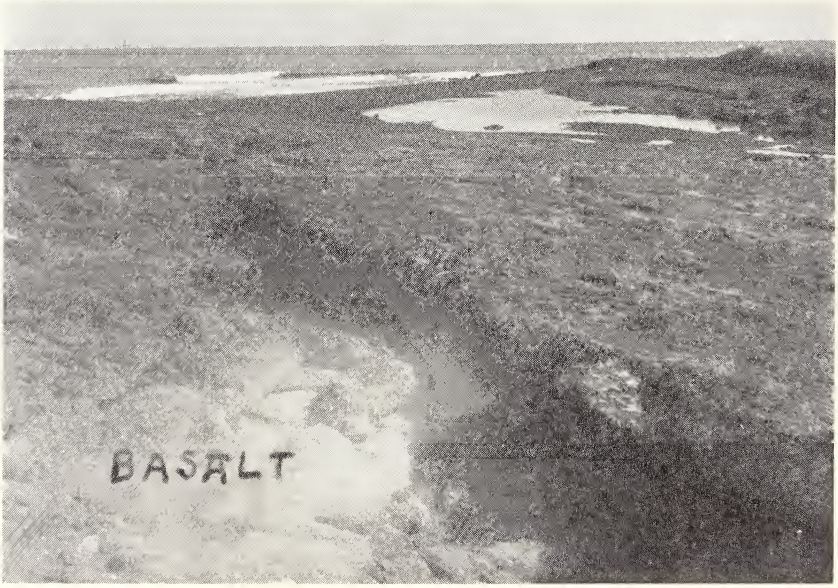

Fig. 11. Remnants of a fossiliferous conglomerate layer some m above the high tide level, Punta del Viento, to the south from Puerto de Cabras.

Photo by H. Hausen 1954

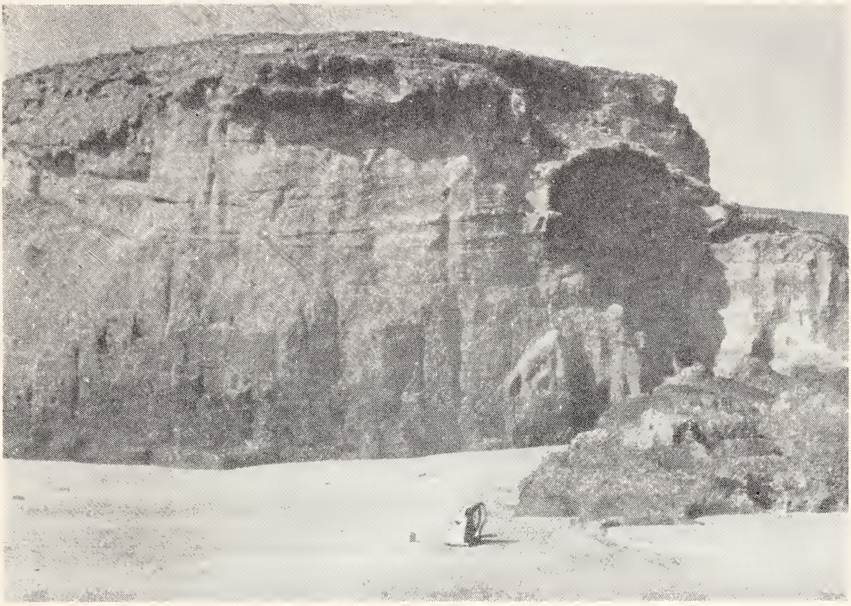

Fig. 12. An abrosion cliff in the friable calcareous sandstone (a hardened dune sand) at Playa Blanca to the south from Puerto de Cabras.
Photo by H. Hausen 1953

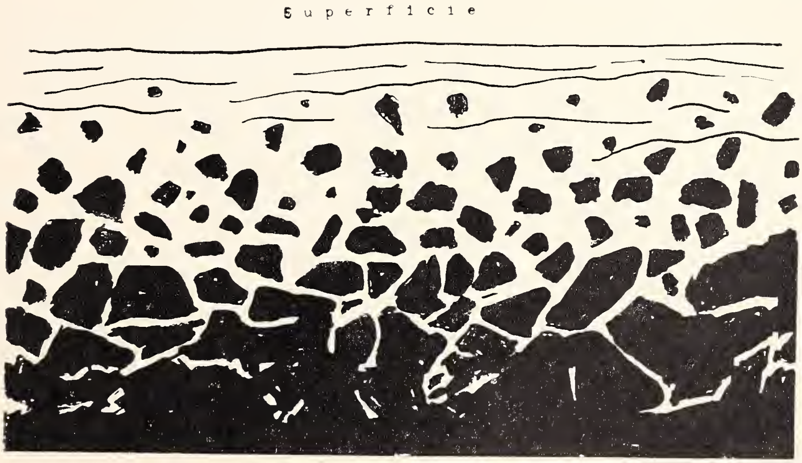

Fig. 13. Idealized section of a travertine covered rock ground illustrating the transition from the rock below through a zone of disseminated rock fragments in the limestone to the pure travertine at the surface. The fragments have been pushed upwards by precipitated lime carbonate; the bicarbonate solution rising from below.

ACTA GEOGRAPHICA 15, N:o 3

CORRELATION OF TREE-RING CHRONOLOGIES OF ALASKA, LABRADOR AND NORTHERN EUROPE

BY

ILMARI HUSTICH

HELSINKI—HELSINGFORS 1956

TILGMANNIN KIRJAPAINO. HELSINKI 1956

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INTRODUCTION

The purpose of this paper is to compare tree-ring series from three different parts of the northern coniferous zone. Interest has been mainly focussed on the following question: do the »minimum years» of the radial growth series coincide over the whole area covered by this study or is there a regular or irregular pattern in the occurrence of the minimum years?

The material is to meagre to allow of a thorough analysis. A preliminary outline of the problem might, however, be of interest to climatologists also, because of the fact that »good» and »bad» years in the growth in thickness of trees to a certain degree reflect the variations of the summer climate in decades and in areas for which we have no meteorological data.

The field research involved in this paper represents part of the author's forest-botanical work in the Labrador Subarctic forests. The work was sponsored by the *Arctic Institute of North America* and the *National Museum of Canada,* for which assistance I here express my sincere gratitude.

Helsingfors, March 1956. *Ilmari Hustich*

I. BIOLOGICAL BACKGROUND

The growth processes of a tree are not straightforward physiological processes. We cannot use the tree-rings as climatic tools comparable to meteorological instruments. The value of tree-ring work has sometimes been overestimated, primarily because of our limited knowledge of the biological and geographical background of tree growth. For a critical compilation of recent work on tree-rings and climate, compare also GLOCK 1955.

The growth in thickness (i.e., radial growth) of coniferous trees near the polar timber line is known to react very sensitively to annual variations of the climate, particularly the summer temperature (compare inter alia EIDE 1926, ERLANDSSON 1936, GIDDINGS 1941, SCHOVE 1954 and H 1948[1]). Because of the short growing season near the timber line the annual variations in the mean July temperature is a good indicator of radial growth variations. In southern and more arid regions, however, the precipitation (or better, the water supply during the growing season) is the limiting factor (compare, inter alia, SCHULMAN 1951 and GLOCK l.c.). Re the differences in growth physiology in regions near the polar timber line and in areas near the semidesert, see also H 1949. Scientists working with tree-rings in areas between the climatically more extreme regions are usually cautious and pessimistic concerning the value of tree-rings as climatic tools (compare DOBBS 1951). The reason is that in these regions we do not have any such clear limiting factors as the temperature in the north or the water supply in the arid regions.

The polar limit of the coniferous zone is not formed by the same species all over the Subarctic. It is true that the trees react habitually very similarly near the timber line but in the details of the growth processes there are, nevertheless, some marked differences.

The map in Figure 1 shows the polar limit of the spruce genus, *Picea*. The location of the material used in this paper is approximately indicated on the same map. The pine genus, *Pinus*, does not enter into the timber line zone in the same degree as the spruce (compare H 1953). Only in northernmost Scandinavia are the timber line and the polar limit of the coniferous zone formed by pine.

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[1]) The expressions H 1948, H 1949, etc. refer to the author's earlier papers.

The dominance of spruce at the timber line all over the Subarctic has led the author to *use only growth series of spruce*. There is, moreover, another reason for not mixing spruce and pine series, in spite of the many long pine series available from Northern Europe, namely that spruce and pine do not react entirely similarly to the annual variations in the summer climate. ORDING (1941, p. 206) compared spruce and pine series from the same forest in Eidsvold, Norway, for the period 1851—1937 and got the correlation coefficient, $r = 0.{23} \pm 0.{10}$.

A large body of Swedish data on tree-rings has recently been published by EKLUND, 1954. From his graphs one can draw the conclusion that the minimum years »shift» more in the pine series, from south to north or from one climatic region to another, than in the spruce series. The same fact appears in MIKOLA's graphs for Finland (1950).

In the series based on growth of pine we also note a »lag» effect more clearly than in the spruce series. The lag effect, as described and illustrated by HUSTICH and ELFVING in 1944, means that the effect caused by an unfavourable summer is »forwarded» to the following growing season. We can thus have minimum years in radial growth of pine even in summers which are not particularly unfavourable for growth in general.

In the spruce series the lag effect does not seem to be as marked as in the pine series. ORDING (l.c.) also assumes that pine is more affected by the last year's climate than spruce.

On the other hand, in the spruce series we probably have a greater cone year effect, as shown by TIRÉN 1935, among others, and EKLUND (l.c.). The amount of cones on a spruce varies from year to year. In years with a great amount of cones the radial growth of the tree is affected. The fruit or cone formation decreases the nutrient supply used for the cambial activity (compare e.g. ANTEVS 1925, TOVSTOLES 1938, GLOCK 1955 and H 1948).

In regions were the climatic conditions are not as severe as in the timber line zone, the negative effect on radial growth of increased cone production in some years is probably greater, because the climate itself in such intermediate regions does not influence the annual variations of growth phenomena to the same degree as in climatically extreme regions. It seems that cone production does not shift the minimum years in spruce as much as the lag effect does in pine series.

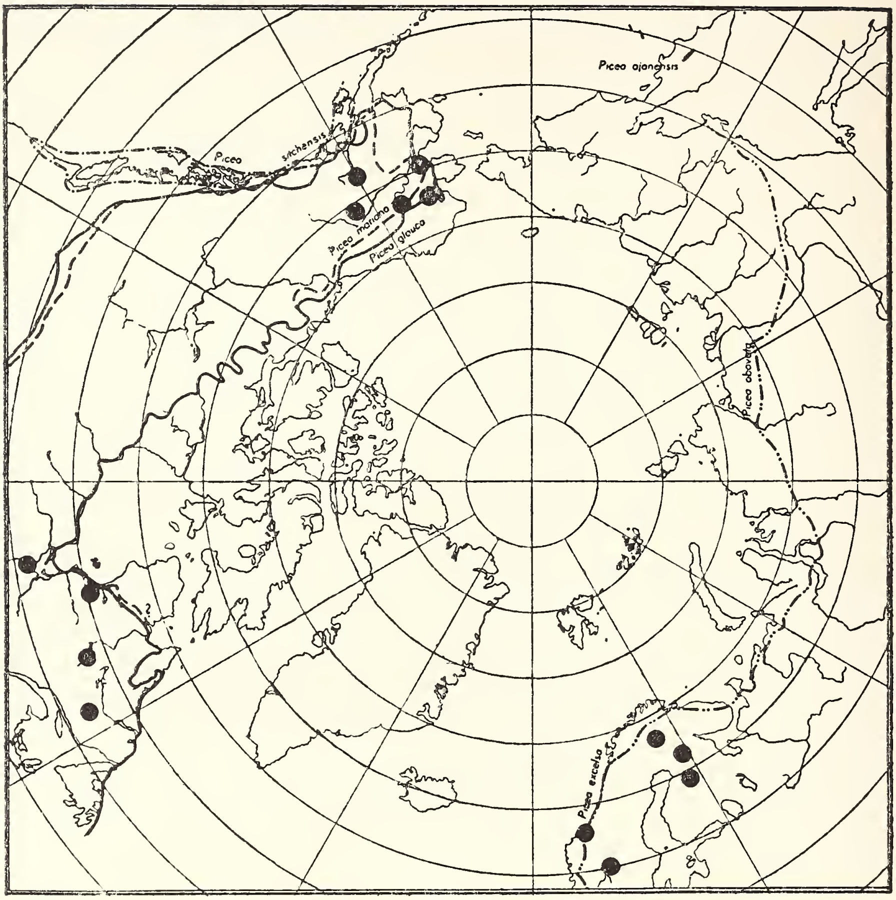

Fig. 1. The localities from which tree-ring data where collected. The boreal limit of spruce is marked on the same map.

Hence in this paper dealing with the teleconnection of tree-ring series, I have used spruce instead of pine series for the following reasons:

1. For dendrochronological purposes it is preferable to use one »type» of tree only. When we cannot use the same species, one should at least use species of the same genus. We can assume that different spruce species react more similarly (note type of needles and cones) to the annual climatic changes than, for instance, spruce and pine.

2. Spruce dominates the Subarctic zone in Europe and North America, whereas the distribution of pine is more irregular (see H 1953).

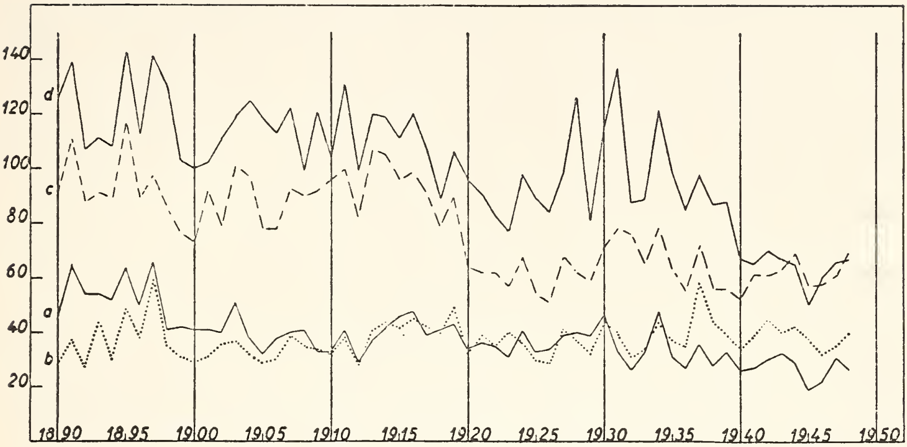

Fig. 2. The annual variations of tree-growth in different forest types approximately illustrated with material from Knob Lake, Labrador. *a* = white spruce lichen forest (3 trees), *b* = white spruce feather moss forest (4 trees), *c* = white spruce bunchberry forest (3 trees) and *d* = rich white spruce forest (4 trees), cf. H 1954, p. 48—50. No smoothing of the age factor has been done. According to microscopical measurements in 0.01 mm.

3. The lag effect in the pine tree-ring series masks the climatic effect more than the varying cone production intensity seems to do in the spruce series.

*

How do the different forest types, e.g. the different moisture conditions on different habitats, influence the tree-ring series? Before we claim any considerable value for the tree-ring series of spruce as a useful climatic tool, we must be sure that the minima in different tree-ring series of spruce occur in the same year over large areas and more or less independently of the different forest types.

In the north, the warm summers are usually also dry summers and the cold summers are usually wet. In dry summers trees in moist and boggy habitats show an increased radial growth because of the increased temperature and the still sufficient water supply in the ground. In drier forest types we might have an occasional decrease in radial growth during very hot and dry summers. In my experience, however, such a decrease in radial growth is not enough to represent a real »minimum year» in the tree-ring series in areas near the timber line, where a drought seriously influencing the growth of mature trees probably never occurs. In wet summers the temperature is usually low, taking the growing season as a whole, causing a minimum year in the growth series on what-ever forest type we are studying.

According to this argument we should have the same annual variation in radial growth on wet and on dry forest types. Experience seems to confirm this, as is illustrated by the following graphs from Labrador, Figure 2. The absolute values for growth in thickness vary on the different forest types. We have fast-growing trees in rich white spruce forests (note observations in H 1950, H 1954), we have slow-growing white spruce in lichen woodlands (l.c.). But the annual variations of radial growth are similar.

Re the differences between the spruce species, compare Figure 3, in which tree-ring series of black spruce, white spruce and balsam fir are compared. There are greater taxonomic differences between black spruce (*Picea mariana* (Mill.) BSP) and white spruce (*P. glauca* (Moench.) Voss.) than between white spruce and Norway spruce (*P. Abies* (L.) Karst. = *P. excelsa* Link.). (No direct comparison between tree-ring series of black or white spruce and Norway spruce from the same area can, of course, be made.)

Only these three spruce species, black and white spruce and Norway spruce, are included in the material described below.

*

A tree grows differently in different periods of its life cycle. Many external factors affect the cycle: a. the constant competition between the trees in a forest, going on underground and between the crowns of the trees, b. a forest fire killing the tree only partly or killing the neighbouring trees, c. a sudden storm felling trees and changing the light conditions, or d. attacks from fungi and insects. All these happenings in the forest leave their marks in the growth of a tree as illustrated by the tree-rings. And to all this is to be added the maturing of the young tree and the decreasing growth of the older tree which goes on as a part of the whole forest cycle, as one tree generation succeeds another.

A forest is never static; there is always something happening. But in the light of all these facts, it seems the more dangerous to accept tree-ring measurements as reliable evidence of the annual variations of the climate itself. If an old depressed tree is suddenly enabled to raise its crown and extend its roots in the most suitable directions, it can be seen to display a slowly increased radial growth, which has nothing to do with the climate. A tree scorched by a forest fire shows a decreased radial growth, but sometimes also a suddenly increased radial growth because the ground is richer in useful nutrients than before the fire. If the light conditions are changed, we can read this in the radial growth. Thus, the radial growth of a tree, illustrated by tree-ring series, affords good evidence of the happenings in the growth of a tree or a forest in

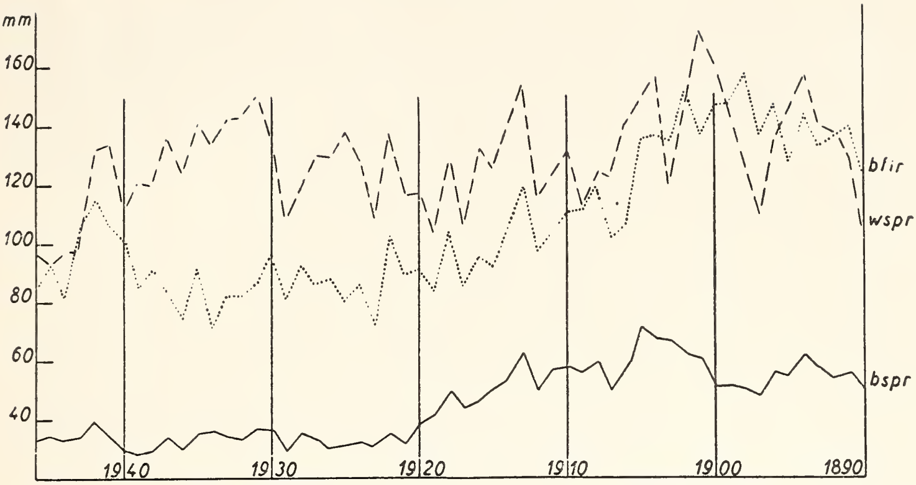

Fig. 3. The annual variations of tree-rings in different tree species, approximately illustrated with material from Moose River, Ontario, cf. H 1955, p. 42. Bspr = black spruce, wspr = white spruce and bfir = balsam fir (Abies balsamea). No smoothing of the age factor has been done. According to microscopical measurements in 0.01 mm.

general. It seemingly registers all the changes in the environment, climatic or edaphic, internal or external, of importance in the life of the tree. This is true and has been stressed above and in many papers. But it does not mean that we know enough of the mechanism of the chain-reaction from cause to visible results and it does not mean that merely by reading the tree-ring evidence we can always separate climatic and edaphic influences or distinguish between internal and external factors.

Among the complications we must include the effects of the normal life cycle of a tree, from a sometimes very slow development in the early years to a rapid increase in growth intensity over an optimum period towards a slow decrease of the growth activity until finally all signs of growth fade out and the tree is dead.

On different habitats the trees attain their optimum growth rate at different times and this means that in different periods the sensitivity to changes in environment and climate is different. The problem arises: how long a period in the life of a single tree should we use for dendrochronological purposes and what correction coefficients must be used to connect the optimum period of growth with the periods with slowly decreasing or increasing growth?

Contrary to general opinion, the trees in the north seldom reach a great age in stands. The oldest ordinary forest tree the author has studied was about 300 years old. Solitary trees on mountain slopes sometimes grow older, but are not according to my opinion very useful for dendrochronological purposes. To obtain growth series going back to 1760, for instance, is in itself a problem in the north. This certainly limits the value of northern dendrochronology. Its use must be restricted to regions from which there are no meteorological observations or where the meteorological series are short.

If we use a series of comparatively old trees, all of the about the same age, we get a distorted picture of the annual growth variations. The earliest years in such a series show an exaggerated growth in thickness with marked annual variations, the later years a slowly decreasing growth with less marked annual differences.

The smoothing of the age factor used by so many scientists, particularly by authors approaching the problem from the forester's angle, is not sufficient. We can, of course, »smooth» the absolute values of growth and eliminate the increased and decreased growth of a tree during different periods, but we cannot smooth or »correct» the varying intensity with which a tree reacts to the annual variations in the climatic environment during the different periods of its life cycle.

The ideal tree-ring series for climatic purposes would probably be a series composed of normally grown trees of several age classes, in order to obtain for each decade or period a sufficient sensitivity to the annual variations i climate. To a certain degree, even though it is not, of course, statistically correct, we meet this requirement by using simple mean values for the radial growth of trees of different age classes in one and the same series. But we generally use more trees for the later decades than for the earlier decades. There are more young and middle-aged trees available for growth analyses than old trees (which also often show signs of fire or other damage). When we try to find minimum years of growth, i.e. narrow tree-rings only, the simple method outlined above can be used, but when looking for longer periods of good or bad years and trying to coordinate these with the trends or waves found by the climatologists, the dendrochronologists must learn to use modern statistical methods.

*

In view of the facts outlined above we cannot expect our series in Table I to be instruments as precise as the different mechanical devices at a meteorological station. However, even the climatologists have trouble with their older series. Røstad writes that several kinds of errors may occur in the earlier meteorological series: »For example, the thermometers were placed on a northern wall exposed to radiations of various kinds, the temperature was observed only once a day and the time of observation varied by hours from day to day, the surroundings of the observational places were seriously altered as shading and sheltering park trees were allowed to grow up, the radiation conditions were altered by development of the surrounding grounds, by newly erected buildings, etc.» (1955, p. 6).

If however, we, limit our scope to looking merely for years with very narrow tree-rings, we can use the tree-ring series in Table I with some confidence re the variations in the macroclimate during the growing season in the regions we are dealing with in this paper. — The Arizona school of dendrochronology sets the greatest importance on »missing rings», i.e., years when the climatic conditions have been as extreme to leave no mark in the radial growth of the tree. In my experience, we do not have such extreme years in the northern forests, because there is always water enough in the ground to allow at least the formation of two or three cell layers (compare MARR 1948 and H 1950).

II. MATERIAL

All the tree-ring series in Table I have been previously published, except the Labrador series *e*. The material is not, however, uniform. The North American series have not been smoothed or corrected whereas the series from Northern Europe have been smoothed according to methods outlined by the authors in their papers mentioned below. As I have stressed in Part I, our interest is concentrated on the occurrence of certain minimum years over wider areas. Therefore the unconformity of the series is no serious obstacle to the analysis.

The *Alaskan* series (A in Table II below) are from GIDDING's paper of 1941. Series Aa is identical with Table 3, p. 92, in GIDDINGS 1941: 10 trees from White Mountains. Series Ab = Table 4, l.c., p. 92: 28 trees from Alaska Range. Series Ac = Table 5, l.c., p. 93: 11 trees from the Noatak area. Series Ad = Table 6, l.c., p. 93: 25 trees from Kouyk, Norton Bay. Series Ae = Table 16, l.c., p. 97: 6 trees from Hunt River. All trees represented in the Alaskan series are presumably white spruce (?). (Re methods used by GIDDINGS, compare l.c., p. 8—10, see also his paper of 1952.)

The *Labrador* material (L in Table II) is of about the same size as the Alaskan material. Series La is from H 1955, p. 42: 31 black spruce trees from Moose River, Ontario. Series Lb = H 1950, p. 78: 6 black spruce trees from Great Whale River, Hudson Bay. Series Lc = H 1954, p. 51: 5 black spruce trees from Knob Lake in Central Labrador. Series Ld = H 1954, p. 51: 15 white

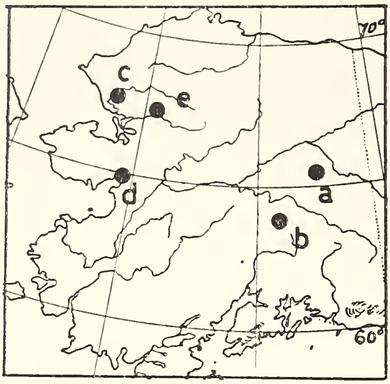

Fig. 4. Location of Giddings' (1941) collections included in Table I, compare p. 11.

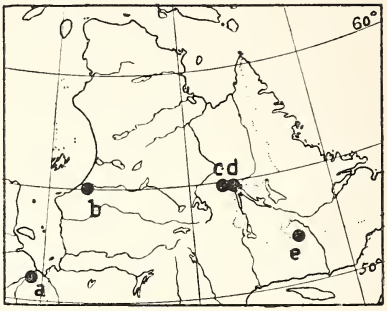

Fig. 5. Location of the author's material from Labrador included in Table I, compare p. 11—12.

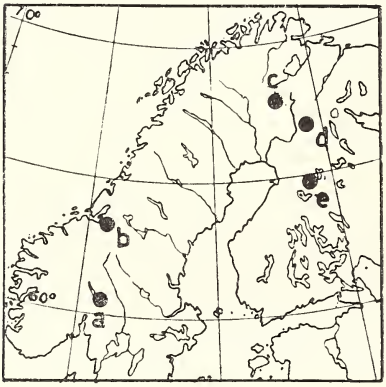

Fig. 6. Location of Ording's (1941 = *a*) Eidem's (1953 = *b*) and Mikola's (1950 = *c*—*e*) material from North-Europe, compare p. 12.

spruce trees from Knob Lake. Series Le = unpublished material of 13 black spruce trees from Mecatina, halfway between Seven Island and Goose Bay, collected in 1952.

The *North-European* material (N) consists of Norway spruce series, partly from the maritime part of the area (Na—Nb), partly from the more continental parts of Finland. Series Na = Ording 1941, p. 182: 30 trees from Eidsvold near Oslo in Norway. Series Nb = Eidem 1953, p. 74: 81 trees from Selby near Trondheim in Norway. Series Nc—Ne are represented as graphs in Mikola 1950; for this purpose Dr. Mikola has kindly allowed me to use the original series, from which the graphs in his paper were drawn. Series Nc = 95 trees from Sompio and Pallastunturi in Finnish Lapland. Series Nd = 72 trees from Savukoski in eastern Finnish Lapland. Series Ne = 139 trees from Kuhmo, Puolanka and Sotkamo in northeastern part of Central Finland.

Re the situation of the localities mentioned above, see Figures 4—6. As can be seen from the maps, the localities are climatically different. The Alaskan and the North-European series are partly from predominantly maritime regions (Ac, Ad, Na, Nb), whereas the Labrador and the Finnish series are from more continental regions.

The tree species represented are black spruce (La—Lc, Le), white spruce Aa—Ae?, Ld) and Norway spruce (Na—Ne).

The North-European series are more reliable than the American series, because the material is bigger. It should therefore be desirable to compare the American series used here with any material available as a control. One is, unfortunately, limited to series collected by the same authors, because the dendrochronological data for Alaska and northeastern Canada are very meagre; compare also Marr 1948.

Table I. Tree ring series from Alaska, Labrador and Northern Europe.

	Alaska					Labrador					Northern Europe				
Year	a	b	c	d	e	a	b	c	d	e	a	b	c	d	e
1946	—	—	—	—	—	33	23	23	42	19	—	—	99	—	—
1945	—	—	—	—	—	34	22	23	36	18	—	—	95	76	76
1944	—	—	—	—	—	33	22	27	47	18	—	—	82	73	80
1943	—	—	—	—	—	34	21	24	44	19	—	—	93	87	94
1942	—	—	—	—	—	39	24	24	49	24	—	—	113	118	108
1941	—	—	—	—	—	35	23	26	47	20	—	—	121	124	114
1940	—	—	34	22	54	29	16	22	43	20	—	—	126	109	139
1939	—	—	36	28	54	29	20	26	51	18	—	—	126	111	113
1938	—	—	30	25	46	30	21	28	51	16	—	—	128	109	109
1937	45	18	39	28	51	34	23	32	58	23	105	120	122	108	104
1936	48	21	32	27	50	30	20	27	46	20	98	140	124	114	128
1935	53	19	34	28	48	35	26	25	52	23	89	110	90	82	94
1934	46	21	24	24	37	36	25	30	65	22	80	100	90	93	110
1933	37	21	22	21	35	34	22	31	50	18	90	105	105	101	108
1932	32	20	27	21	43	34	26	27	53	20	103	95	102	98	100
1931	35	18	22	20	42	37	25	31	63	20	96	75	83	70	78
1930	38	18	27	22	45	37	23	37	62	20	133	130	105	94	95
1929	41	17	31	24	43	29	22	31	44	19	117	110	112	105	110
1928	21	15	28	25	39	35	26	34	53	19	89	55	84	85	72
1927	32	19	34	31	34	33	25	30	55	22	97	110	145	130	117
1926	32	15	31	29	35	30	17	26	40	17	102	110	123	107	117
1925	31	16	34	29	33	31	20	31	49	21	98	95	137	98	100
1924	23	15	24	25	24	32	25	32	56	22	94	85	140	115	120
1923	39	19	27	24	35	31	25	24	46	21	92	65	115	104	111
1922	22	15	14	16	25	35	22	28	49	21	99	90	133	147	129

Table I. Tree ring series from Alaska, Labrador and Northern Europe.

	Alaska					Labrador					Northern Europe				
Year	a	b	c	d	e	a	b	c	d	e	a	b	c	d	e
1921	30	19	24	26	47	32	24	26	52	23	95	60	105	118	102
1920	34	19	26	29	38	38	20	27	52	25	99	80	92	96	80
1919	19	10	19	20	29	42	24	31	66	25	109	105	112	131	114
1918	36	17	28	23	35	49	23	32	59	23	86	70	80	96	73
1917	32	15	20	20	36	44	26	33	69	23	95	85	87	107	115
1916	28	15	20	23	38	46	22	34	76	26	101	85	95	121	117
1915	34	14	18	26	31	50	20	29	69	25	103	80	92	119	118
1914	29	14	24	30	45	53	24	33	70	26	135	105	91	106	106
1913	37	17	25	27	39	62	22	28	67	28	135	80	78	84	93
1912	19	10	15	19	25	49	20	25	58	27	135	105	75	83	95
1911	35	16	19	32	36	57	30	30	75	30	127	95	94	99	104
1910	29	15	19	25	47	57	25	27	65	25	152	110	81	76	77
1909	34	15	17	25	36	56	22	26	66	25	116	95	101	90	88
1908	36	16	24	30	44	59	22	30	66	30	131	115	95	82	82
1907	37	15	28	33	44	49	23	31	68	27	98	105	103	100	75
1906	40	17	28	32	45	57	26	30	62	28	106	110	91	78	74
1905	33	15	22	31	40	71	25	26	59	28	114	95	109	99	92
1904	28	12	21	33	36	67	24	31	68	32	96	80	96	90	73
1903	37	18	23	34	30	66	25	38	68	30	111	85	83	80	87
1902	33	18	24	36	33	62	25	35	60	26	78	70	72	76	76
1901	38	18	18	28	26	60	24	34	65	30	137	125	120	119	123
1900	26	15	20	34	34	51	24	30	57	31	125	115	86	106	127
1899	20	12	13	16	28	51	23	34	61	30	149	105	91	117	111
1898	32	16	21	34	43	50	23	36	68	30	119	95	94	108	109
1897	33	18	28	32	37	47	26	45	79	39	105	100	80	91	91
1896	35	17	19	29	46	55	26	41	71	36	106	90	96	121	109
1895	36	17	24	34	50	54	27	40	90	43	103	95	88	107	106
1894	22	17	28	36	57	61	24	35	71	36	99	75	81	82	93
1893	27	17	27	36	41	57	26	36	71	38	108	100	107	102	123
1892	23	17	22	38	37	54	23	31	67	42	99	75	103	98	111
1891	35	18	23	37	48	55	27	36	86	40	101	105	108	95	95
1890	34	19	28	37	54	49	26	30	70	38	95	100	112	111	127
1889	24	13	26	31	52	55	22	32	77	40	107	140	117	121	131
1888	36	17	27	28	42	52	27	35	89	42	116	100	92	90	109
1887	25	13	32	27	56	51	36	35	82	41	127	85	82	75	95
1886	33	18	24	27	42	60	33	32	85	40	152	110	123	107	113
1885	36	14	23	30	52	63	33	35	77	39	130	85	115	104	126
1884	38	19	22	30	68	57	36	32	74	39	140	105	108	102	107
1883	31	13	23	27	46	62	34	37	76	40	107	125	114	115	98
1882	44	18	31	34	53	66	36	26	66	45	120	145	88	88	89
1881	39	19	33	33	54	67	36	35	90	40	82	100	84	93	95
1880	43	17	31	33	60	66	35	36	87	47	92	100	64	81	84
1879	44	14	26	32	58	68	30	36	82	45	79	115	71	87	105

	Alaska					Labrador					Northern Europe				
Year	a	b	c	d	e	a	b	c	d	e	a	b	c	d	e
1878	38	18	34	39	55	66	34	33	83	47	101	115	86	102	109
1877	40	20	36	39	73	58	28	29	71	45	81	105	90	116	100
1876	41	21	22	37	56	56	26	26	70	50	99	105	85	109	89
1875	47	20	35	36	41	61	27	33	88	49	135	120	78	112	93
1874	43	18	29	36	52	61	38	32	79	46	127	100	47	81	76
1873	38	17	28	30	50	65	41	30	84	46	94	115	77	111	97
1872	30	13	32	29	42	65	35	28	70	46	129	145	92	112	84
1871	49	17	26	32	43	74	30	28	63	47	112	100	84	91	61
1870	49	16	31	33	36	56	37	30	65	46	131	110	101	100	53
1869	42	14	30	35	40	57	26	31	64	43	96	75	101	93	52
1868	45	16	31	35	33	53	34	28	73	44	116	105	130	121	99
1867	58	20	38	33	47	65	33	32	69	43	90	75	127	116	97
1866	43	17	30	28	42	67	28	30	65	45	115	95	122	105	118
1865	50	21	36	32	48	63	36	32	63	46	108	80	126	114	99
1864	42	21	28	36	39	62	38	33	63	46	108	95	122	107	121
1863	46	22	34	32	42	66	34	38	66	49	118	105	120	105	115
1862	52	25	28	28	43	78	32	31	61	49	122	115	106	103	107
1861	58	23	29	31	50	77	38	31	61	43	131	135	105	99	113
1860	58	25	27	39	55	59	33	28	63	38	122	130	97	103	106
1859	73	27	27	34	48	64	32	29	66	36	108	100	68	71	77
1858	62	24	37	44	43	60	30	31	82	41	96	105	78	87	87
1857	66	25	45	52	59	51	26	22	61	36	107	90	63	74	89
1856	38	16	29	35	39	67	30	27	82	41	97	75	75	101	99
1855	57	24	44	46	53	70	31	31	71	40	98	95	64	97	96
1854	59	24	37	37	48	62	32	36	93	46	120	100	69	105	105
1853	52	22	39	41	51	68	33	32	75	46	97	80	114	122	113
1852	40	20	31	36	47	65	33	28	68	43	107	105	120	123	109
1851	49	19	21	30	49	81	37	36	79	42	77	55	104	112	102
1850	34	19	24	35	50	74	37	37	78	41	86	85	99	102	115
1849	42	20	27	34	52	54	33	29	74	44	73	65	115	105	109
1848	64	29	25	34	49	72	30	24	72	40	96	95	85	90	102
1847	47	23	26	37	49	52	25	27	79	42	66	105	96	111	96
1846	43	20	19	30	52	56	30	26	77	54	83	105	89	102	102
1845	35	19	18	28	52	62	32	24	74	41	98	105	93	111	102
1844	53	28	24	36	45	56	29	21	70	39	104	75	81	83	97
1843	47	20	23	31	39	59	24	22	70	41	97	115	67	74	76
1842	54	20	22	27	46	58	30	22	72	37	111	110	77	85	101
1841	49	21	21	26	36	58	38	21	64	27	90	90	94	128	138
1840	47	19	19	29	40	70	32	19	79	33	80	80	86	133	119
1839	35	16	23	30	45	87	35	17	76	33	84	105	78	104	116
1838	52	23	22	35	59	64	29	14	60	27	74	100	112	124	113
1837	50	18	23	36	49	76	33	13	71	30	79	95	79	85	87
1836	45	20	17	30	34	61	37	15	65	30	88	90	97	84	80
1835	34	16	20	36	37	74	29	15	88	31	74	85	93	96	87

Table I. Tree ring series from Alaska, Labrador and Northern Europe.

	Alaska					Labrador					Northern Europe				
Year	a	b	c	d	e	a	b	c	d	e	a	b	c	d	e
1834	41	15	19	33	45	59	43	28	81	35	105	100	92	102	100
1833	51	19	21	35	51	71	44	26	77	42	107	100	110	125	111
1832	58	21	22	37	61	71	43	23	81	41	79	80	100	108	98
1831	42	19	25	37	62	67	43	24	95	37	105	120	143	152	127
1830	40	17	22	37	78	78	52	28	92	33	80	105	121	146	127
1829	54	20	23	32	62	71	48	24	102	33	100	125	162	179	141
1828	61	19	22	28	48	71	52	26	90	37	105	140	144	146	137
1827	38	14	18	28	46	66	37	28	91	35	91	120	141	165	142
1826	28	13	17	22	38	62	46	31	80	40	93	140	149	169	143
1825	32	17	22	30	60	77	40	28	71	33	83	85	109	107	120
1824	38	18	23	33	52	70	31	26	88	34	94	110	101	97	99
1823	40	19	29	27	43	77	47	25	106	38	90	105	90	77	99
1822	31	14	15	20	38	85	43	33	96	39	83	105	55	48	73
1821	33	16	25	29	50	118	42	40	93	39	79	50	41	35	52
1820	35	17	32	31	45	112	41	28	93	36	66	110	62	45	66
1819	59	15	26	29	45	100	38	21	76	26	111	150	—	33	82
1818	45	15	27	28	51	105	40	29	86	38	110	140	—	30	90
1817	42	14	28	28	58	79	42	29	57	—	107	130	—	59	88
1816	41	15	18	21	44	58	32	23	70	—	102	135	—	98	96
1815	41	15	23	24	46	45	35	30	87	—	93	105	—	77	86
1814	50	16	24	27	43	52	31	33	92	—	99	120	—	59	100
1813	55	19	24	28	44	60	34	33	98	—	95	120	—	64	90
1812	45	21	26	29	46	42	44	31	94	—	91	100	—	76	87
1811	54	24	26	28	58	34	40	35	103	—	102	135	—	100	102
1810	49	24	26	28	60	27	40	36	105	—	98	110	—	101	96
1809	40	23	23	25	48	41	40	30	102	—	77	105	—	112	114
1808	50	26	35	34	58	41	47	30	116	—	88	135	—	131	107
1807	39	19	26	29	55	45	58	26	112	—	81	105	—	131	80
1806	45	24	30	29	51	56	51	27	105	—	73	70	—	71	69
1805	45	23	27	30	50	52	47	36	106	—	69	80	—	132	100
1804	41	21	22	31	43	45	50	31	109	—	85	105	—	124	97
1803	54	26	27	38	65	50	52	27	107	—	72	90	—	113	115
1802	53	28	30	43	55	59	69	35	122	—	79	95	—	122	114
1801	54	29	29	40	46	54	52	35	97	—	99	125	—	131	129
1800	48	23	33	37	42	48	49	34	94	—	93	90	—	140	119

To the series 1800—1946 in Table I I have added as a supplement data concerning tree-ring growth in 1760—1800, appearing in Table III. The scantiness of my Labrador material for earlier years does not allow us to go beyond 1760.

Of the Alaskan series used in Table I Aa—Ad go only to 1800, whereas series Ae starts in 1720, but comprises only 6 trees. For the period 1760—1800 the following tables by GIDDINGS (1941) have been used: 32 (Haycock, 2—3 trees), 31 (Interior Timber line, 3 trees), 25 (Stephens village, 5 trees) and 24 (Fort Yukon, 2—6 trees) to give an approximate idea of the occurrence of minima before 1800; see Table III.

Of the Labrador series in Table I, La starts 1799 and Le 1818. The Great Whale series in Table I, Lb, starts 1764 (only 3 trees included for the earliest years). The Lc-series can also be used for the period 1760—1800, but only 3 trees can be included. Of the Knob Lake white spruce series, Ld, 4 trees are old enough to be used for the continuation of the series down to 1760. In Table III to enlarge the very meagre material of old trees available so far from Labrador I have also used MARR's »master charts» (1948) from Great Whale River (8 trees) and Richmond Gulf (? trees), some 90 miles N. of Great Whale River.

Of the North-European series, EIDEM's and ORDING's long series have also been used for the period 1760—1800. Of MIKOLA's series only Ne goes down to 1771; the two others to 1820 and 1800 respectively. Thus the supplementary data for Northern Europe in Table III are based on only two series (from 1771 on three series). There are several spruce series available for Norway, but not from Finland. Re a comparison with SCHOVE 1954, compare p. 22—25.

*

In this paper no climatological data are given. Such material is simply not available for any longer periods from Alaska or Labrador. Also for an area like Labrador, rather near to the early populated St. Lawrence valley, there are few reliable climatic data from earlier decades. According to information, kindly forwarded by Professor KENNETH HARE, Montreal, there are few data from Moose River (1890, 1897—1908, 1916—), Clarke City near Seven Islands (1907—) and Mistassini Post (1914—). And even these series are broken and unhomogeneous for one reason or another. In such regions surely dendrochronological material must be of some help to climatologists dealing with the problem of the annual variation in the summer climate of the Subarctic.

III. ATTEMPTED ANALYSIS.

Even a preliminary scrutiny of the material in Table I shows that the minimum years in radial growth seldom occur simultanously over the whole region covered in this paper.

When comparing two tree-ring series, we can to a certain extent use ordinary formulae for the calculation of the correlation coefficients. The size of the correlation decreases as the distance between the localities increases and particularly when moving from one climatic region to another.[1] Particularly marked discrepancies are noted in areas where the topography is rugged and the local climate, as in Norway, changes rapidly from one place to another. Similar discrepancies must also occur in Alaska, whereas in Labrador and Finland, for instance, we have the same annual variations over large areas.

The distances between Northern Europe, Alaska and Labrador are so great that no positive correlation can be expected. Instead, we obtain small, entirely insignificant negative coefficients when comparing, for instance, series from northern Finland and the more maritime Alaskan region, compare H 1948, p. 43. Thus the usual correlation calculations are of no interest in the intercontinental tree-ring correlation attempted here. But over wide areas of the same continent and within the same climatic regions, we can at least find some definite minimum years occurring simultaneously. And if we move over still larger areas and from one continent to another, the spectacular minimum years in radial growth do not occur in the same growing season, but start shifting from one year to another.

But why not also take into account the years with exceptional growth in thickness? If the summer temperature in general is low we have decreased radial growth of spruce in every type of forest in the north. If the summer temperature is higher than normal we might have in some drier forest types a limited growth in thickness (cf. above). Empirically, it seems that in tree-ring series from the north the maximum years of growth in thickness do not coincide as closely as the minimum years of growth, the maxima often shifting more easily.

For obvious reasons, the author has omitted correlation analyses and limited his analysis of the tree-ring series in Table I to the very simple method of indicating the minimum years only. This method is in practice more or less the same as the »skeleton method» used for dating purposes by the Arizona school of dendrochronology. But my aim is not, of course, dendrochronological dating, but only to correlate some series of tree-ring data of possible climatological interest.

---

[1] As an example I will quote some correlation coefficients (1890—1939) for pine series from northern Scandinavia (from H 1948, p. 43 and 47): Enare/Utsjoki, distance about 60 miles, r = 0.89 ± 0.04, Karesuando/Utsjoki, distance about 150 miles, r = 0.74 ± 0.07, Abisko/Utsjoki, distance about 180 miles, r = 0.83 ± 0.05, Sörfold, Norway/Utsjoki, distance about 300 miles, r = 0.63 ± 0.09. Compare, however, also Ording 1941 and Mikola 1950.

For a study of the occurrence of minima in the tree-ring series in Table I the series have been simplified as follows: For each group in Table I, Alaska, Labrador and Northern Europe, the five series have been compared with each other from year to year using an arbitrary point system: If in all five series in one group a certain year appears with a minimum of growth in thickness compared with the preceding and following years, the year in question is valued *5*. If in four of the five series a certain year appears as a minimum year the year in question is valued *4*, etc. Of course, two years valued with *5* need not have the same narrowness of radial growth: the main thing is that both years appear as minimum years compared with the preceding and following years. By this simple method, however, we have made the relative annual differences in each series comparable with each other and have thus at least got an idea of which minimum years are common to the regions in question.

The primitive points system outlined above is illustrated by Table II. The minimum years marked in this Table are thus years during which — in spite of habitat, forest type, age of the trees included in the series, cone formation intensity, etc. — there has been a decreased growth in thickness.

Table III comprises only the more prominent minimum years valued with *3—5* points in each region.

At first sight it is difficult to find any regularity in the occurrence of years with narrow tree-rings. *Using a such »master chart» it would be extremely difficult to do any cross-dating of tree-rings over the Atlantic or over the whole northern part of North America. Tables I and II actually present a very strong argument against any such attempts.*

It is of interest to compare the series in Tables I and II which each other: Alaska with Labrador, Labrador with Northern Europe, Northern Europe with Alaska, looking for minimum years which occur simultaneously in at least two areas.

Comparing Alaska and Labrador, we find that the following years are minima in both series: 1915, 1912, 1892, 1869, 1800, 1798, 1794 and 1765.

In the Labrador and North-European series we have the following identical minima: 1923, 1918, 1912, 1894, 1892, 1871, 1869, 1800, 1786, 1774 and 1767.

In the series from Alaska and Northern Europe the following minima are identical: 1931, 1928, 1912, 1904, 1892, 1887, 1869, 1851, 1800 and 1780.

Only the following years appear simultaneously as minima in Alaska, Labrador and Northern Europe: 1912, 1892, 1869 and 1800.

Re the different value of the material from the period 1760—1800 and from 1800—1946, compare p. 16—17.

Table II. Minimum years in the Alaska, Labrador and Northern Europe series in Table I. For explanation of the numbers 1—5, see p. 19.

	A	L	N		A	L	N		A	L	N
1946	—	—	—	1903	—	—	—	1860	—	2	—
1945	—	2	1	1902	—	2	5	1859	1	1	4
1944	—	1	2	1901	3	—	—	1858	2	—	1
1943	—	2	—	1900	—	2	3	1857	—	5	2
1942	—	—	—	1899	5	—	—	1856	5	—	2
1941	—	—	—	1898	—	—	1	1855	—	2	2
1940	2	4	1	1897	1	1	4	1854	3	1	—
1939	—	—	—	1896	2	2	1	1853	—	—	2
1938	3	1	—	1895	—	—	—	1852	2	3	—
1937	—	—	1	1894	1	3	5	1851	3	—	3
1936	1	4	—	1893	—	—	—	1850	1	1	2
1935	1	1	3	1892	3	4	3	1849	—	1	2
1934	—	—	2	1891	—	—	2	1848	2	3	2
1933	2	3	—	1890	—	4	2	1847	—	2	2
1932	1	1	—	1889	3	1	—	1846	—	—	—
1931	3	—	5	1888	1	—	—	1845	4	—	—
1930	—	—	—	1887	3	2	4	1844	—	4	1
1929	—	5	—	1886	1	1	—	1843	2	1	4
1928	3	—	5	1885	1	—	2	1842	—	—	—
1927	1	—	—	1884	1	4	1	1841	2	3	—
1926	1	5	—	1883	4	1	1	1840	1	1	2
1925	—	—	1	1882	—	2	2	1839	2	—	2
1924	4	—	—	1881	1	1	1	1838	—	4	1
1923	—	4	5	1880	1	1	4	1837	1	1	1
1922	5	1	—	1879	2	1	1	1836	3	2	2
1921	—	2	1	1878	3	—	—	1835	1	1	2
1920	—	1	3	1877	—	—	2	1834	3	1	1
1919	5	—	—	1876	1	4	1	1833	—	1	—
1918	—	3	5	1875	1	—	—	1832	—	1	5
1917	1	1	—	1874	—	1	4	1831	—	1	—
1916	1	—	—	1873	1	—	2	1830	3	1	2
1915	3	4	1	1872	4	—	—	1829	—	2	—
1914	1	—	—	1871	1	3	4	1828	—	1	1
1913	—	—	1	1870	1	1	—	1827	—	2	—
1912	5	5	3	1869	3	3	4	1826	5	1	—
1911	—	—	2	1868	1	2	—	1825	—	2	2
1910	1	1	3	1867	—	1	3	1824	—	2	—
1909	3	2	2	1866	5	2	1	1823	—	1	—
1908	—	—	—	1865	—	1	3	1822	5	—	—
1907	1	2	3	1864	4	1	—	1821	—	—	4
1906	—	—	2	1863	—	—	—	1820	—	—	—
1905	—	2	—	1862	1	1	—	1819	1	5	—
1904	3	1	3	1861	1	—	—	1818	—	—	1

	A	L	N		A	L	N		A	L	N
1817	1	1	—	1810	—	1	—	1804	4	1	—
1816	3	2	—	1809	5	1	2	1803	—	2	1
1815	—	1	3	1808	—	—	—	1802	—	—	—
1814	1	1	1	1807	3	1	—	1801	1	—	—
1813	—	—	—	1806	—	1	3	1800	4	4	3
1812	1	2	1	1805	—	1	—				

If we peruse Table II more carefully we find a certain, to some extent almost regular, shift of the minima. If, for instance, 1922 was a minimum year in Alaska, we find that 1923 was a minimum year in Labrador, etc. Below I have listed such cases in Tables II and III. An interesting feature is that the »shifting» of minimum years, as understood above, is almost nil between Labrador and Northern Europe. Here we have to deal with general climatic laws regulating the atmospheric circulation in the Arctic and Subarctic.

In the following cases a minimum year(*n*) in radial growth of spruce in Alaska is followed by a minimum year of radial growth of spruce in the following year $(n + 1)$ in Labrador: 1928 (in Alaska narrow tree-rings) — 1929 (narrow tree-rings in Labrador), 1889—90, 1883—84, 1856—57, 1851—52, 1782—83, 1773—74. From Labrador to Northern Europe we have a similar shifting in 1943—44 (?) and 1783—84. A minimum year in radial growth in Alaska is followed by a minimum the following year in Northern Europe: 1922—23, 1919—20, 1909—10, 1901—02, 1899—1900, 1866—67, 1864—65, 1773—74.

An opposite shifting of the minimum years is prominent in some decades: from Labrador to Alaska, 1923—24, 1918—19, 1871—72, 1844—45, 1772—73, from Northern Europe to Labrador, 1944—45 (?), 1935—36, 1928—29, 1851—52, 1764—65, and from Northern Europe to Alaska, 1923—24, 1918—19, 1900—01, 1871—72, 1865—66, 1821—22, 1815—16, 1806—07, 1764—65.

Of course, all this may well prove to be a pointless play with numbers and years. However, one fact at least seems to be quite clear: very rarely do we find a cold summer occurring simultaneously all over the whole Subarctic.

The correlation between cold summers and narrow tree-rings in the northern forests is fairly well established. There must be some kind of correlation between the general climatic circulation and the tree-ring series, now and in earlier decades and centuries as well. Thus, simple tree-ring data could be useful for outlining these general laws for the circulation with an additional material reaching beyond the meteorological series.

One way towards a connection of climate and tree-rings is pointed out by Schove in his important paper of 1954. He correlated a tree-ring index series with the pressure situation of the summers: »The temperature anomaly

Table III. Minimum years in radial growth of spruce in Alaska, Labrador and Northern Europe.

Alaska = A	Labrador = L	N-Europe		A/L[1]	L/N[2]	N/A[3]	Temperature period *(Willett)*
		= N	*(Schove)*				
	1945						
		1944					warm
	1943						
	1940						
1938							
	1936						
		1935	1935				
	1933						warm
1931		1931	1931			1931	
	1929						
1928		1928	1928			1928	
	1926						
1924							warm
	1923	1923	1923		1923		
1922							
			1921				
		1920					
1919							
	1918	1918	1918		1918		
1915	1915			1915			
			1914				warm
1912	1912	1912		1912	1912	1912	
			1911				
		1910					
1909			1909				
		1907					
1904		1904	1904			1904	
		1902					cold
1901							
		1900	1900				
1899							
		1897					
	1894	1894			1894		cold
1892	1892	1892	1892	1892	1892	1892	

[1] Minimum years occurring simultaneously in Alaska and Labrador, [2] in Labrador and Northern Europe, [3] in Northern Europe and Alaska.

Alaska = A	Labrador = L	N-Europe		A/L	L/N	N/A	Temperature period
		= N	(*Schove*)				
1889			1888				
1887		1887				1887	
	1884						cold
1883			1881				
		1880					
1878							
	1876		(1876)				
		1874	1874				
1872							cold
	1871	1871			1871		
1869	1869	1869	1869	1869	1869	1869	
		1867	1867				
1866							warm
		1865	1865				
1864							
		1859	(1859)				
	1857		1857				
1856							
1854							warm
	1852						
1851		1851	(1851)			1851	
			1850				
			(1849)				
	1848						
1845							
	1844		1844				cold
		1843					
	1841		1840				
	1838						
1836			1837				
1834							cold
		1832					
1830							
1826							
			1825				warm
1822							
		1821	1821				

Alaska = A	Labrador = L	N-Europe		A/L	L/N	N/A	Temperature period
		= N	(*Schove*)				
	1819						
1816							cold
		1815	1815				
			1812				
1809							
1807							
		1806	1806				cold
1804							
			1803				
1800	1800	1800	1800	1800	1800	1800	
1798	1798			1798			
1794	1794			1794			warm
		1790	1790				
1788							
	1786	1786	1786		1786		
		1784	1784				warm
	1783						
1782							
1780		1780				1780	
1778							
1776							
	1774	1774			1774		warm
1773							
	1772						
	1770						
			1769				
	1767	1767			1767		warm
1765	1765		1765	1765			
		1764	1764				

of a particular growing season is often similar over a wide area from the Faroes to Finland. Moreover, it can be correlated with the pressure difference between those two regions — more specifically with the extent and strength in North Scandinavia of the »summer monsoon». — — — Almost invariably the years when tree-rings were narrower than 0.75 mm were years of northerly wind excess in North Scandinavia» (l.c., p. 44—45).

BRIER compared earlier (1948) the departures of mean annual hemisphere pressure from normal for a 41-year period (1899—1939) with the author's tree-ring series (pine) from northernmost Finland: »Years with high mean pressure tend to be years with low mean temperatures in northern latitudes» (BRIER l.c.).

In Table III I have included SCHOVE's tree-ring index series. There are some discrepancies between the minimum years occurring in SCHOVE's series and in Table II, probably mainly owing to the fact that SCHOVE operates with a mixed pine and spruce series (cf. the author's comments on p. 5). In certain periods the minima appear simultaneously: 1923—35, 1859—69, 1784—1806, in others there are differences, but the really prominent minimum years seem to be the same.

*

The author limits his purpose to the teleconnection of three groups of tree-ring series and therefore leaves the interpretation of the connection between climate and tree-rings, and between the general laws regulating the atmospheric circulation and the shifting of minimum years to the climatologists. However, I still believe a tree-ring material of this kind is of some help, particularly if we are modest enough to interpret it with due understanding of its biological and geographical background.

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ON THE PORPHYROBLASTIC GRANITE OF CENTRAL SIERRA LEONE

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ABSTRACT

In the old pre-Cambrium, granitic rocks containing large insets of feldspars and quartz are well known, and often they form large massifs. Particularly in West Africa, with an especial reference to Sierra Leone, the hills composed of such granites are usually devoid of all vegetation excepting grass, and they are of a form of tors or inselbergs, and the porphyroblastic granite of such hills is often penetrated by veins and dykes of latekinematic granite, and embraced by gneiss.

In the present paper, the origin of such granites, and particularly of the mentioned tor-like hills, will be discussed as well as the formation of the mentioned morphological features as related to their geology.

In conclusion, the emplacement of the porphyroblastic granites will be attributed to the doming, and the most recent morphological features of such hills, to the primary tectonics of the rocks, and to their geological evolution.

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INTRODUCTION

Granitic rocks which contain large, usually pinkish porphyroblasts of feldspars and white ones of quartz, are well known in all pre-Cambrian regions. In Finland they occupy large areas in the central part of the country, and they have a large distribution in West Africa, as well.

These rocks have been usually called »porphyritic», in spite of that, that according to our present knowledge, their insets are not phenocrysts but porphyroblasts; hence they are formed later than the matrix. Consequently, they should not be termed »porphyritic» but »porphyroblastic», and the latter name will exclusively be used in the present paper, particularly dealing with West African porphyroblastic granites.

In Finland, the porphyroblastic granites usually form ridges or hills, which are characterized by steep slopes, abundant and regular jointing, and by poor and simple vegetation.

Otherwise, in Finland, the sceneries of the areas occupied by the porphyroblastic granite, do not differ essentially from the regions occupied by ridges and hills composed of other granitic rocks or gneiss.

In West Africa, on the contrary, the hills composed of porphyroblastic granites may be rather typical of the sceneries of certain areas. Virtually one can tell there about the landscape of porphyroblastic granite, which there forms very steep sloped, rounded hills, elevating as rocky pinnacles much over the level of their environs. Being covered by short grass and devoid of any other vegetation, they markedly differ also on this basis from the smooth and low surroundings which mostly are covered by a small bush or by a tall grass of the savannah areas.

Such forms closely resemble the inselbergs, or the tors described from England, where they are interpreted to be a mass of rock left in an elevated position by the decay of surrounding parts.

The inselbergs and tor-like structures composed of the porphyroblastic granite, have been described from several areas of West-Africa, and here may only be referred to Junner (1940), who desribed such tors from the Gold Coast, and to similar descriptions of Bolgarsky (1950) from the French Ivory Coast.

PORPHYROBLASTIC GRANITES IN CENTRAL SIERRA LEONE

The western part of Sierra Leone is comparatively flat, but in its central part, a schistbelt composed of the Sula Mountains and the Kangari Hills, runs across the country, from south to north, and on eastern side of this schistbelt

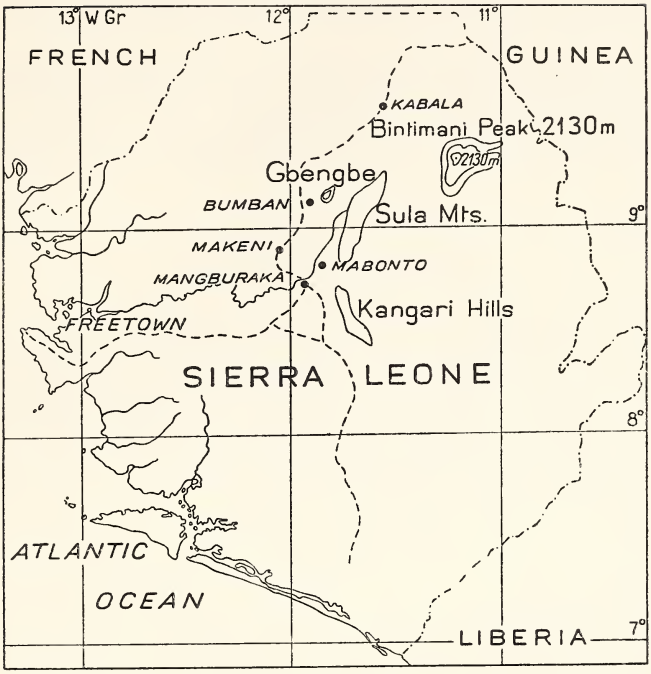

Fig. 1. Location of the described areas.

the country is hilly. There is also, near the boundary against French Guinea (see map Fig. 1) the highest point of Sierra Leone, the Bintimani Peak, which elevates 2 130 m above sea level. Along the western side of the schistbelt, and at its north end, the porphyroblastic granites are videly distributed, and there, the surrounding area being mostly less than 150 m a.s.l., the above-mentioned tor-like structures are especially striking.

On both sides of the motor road between Makeni and Kabala the strange appearance of such tors or inselbergs is drawing attention both of the geologists and geographers, and also of laymen. The rounded hills rise 100 to 200 m or even more above the main level of the area. Their steep slopes are usually black and smooth, and their tops are alike rounded, flat skull-caps, covered by short green grass. Some of hills are splitted into laths (Fig. 7), which, being hardly more than 100 by 50 m at the base, rise upwards 100 to 150 m. Often they are inclined in such an amount, that they resemble the famous tower of Pisa. And all hills are embraced by small bush including some larger single trees. On the hill slopes there are no rice farms, which are typical of the hills of the Sula Mountains in the vicinity. There are no huts on these hills, and the villages are restricted to the river valleys between the hills composed of porphyroblastic granite.

Similar landscapes occur to the south and southwest of Mabonto (map Fig. 1), as well. On both sides of the Kangari Hills, however, the tor-like structures are less conspicuous, but still typical of the porphyroblastic granite.

THE GBENGBE HILLS

The most striking landscape of the areas composed of the porphyroblastic granite is that of the Gbengbe Hills, which situate 35 km NE of Makeni, E of Bumban (Fig. 2 and 5). The area of the Gbengbe Hills is only 200 km², but these mountains form a definite morphological unit within the Central Sierra Leone. The highest point of the Gbengbe Hills is 855 m a.s.l., and almost

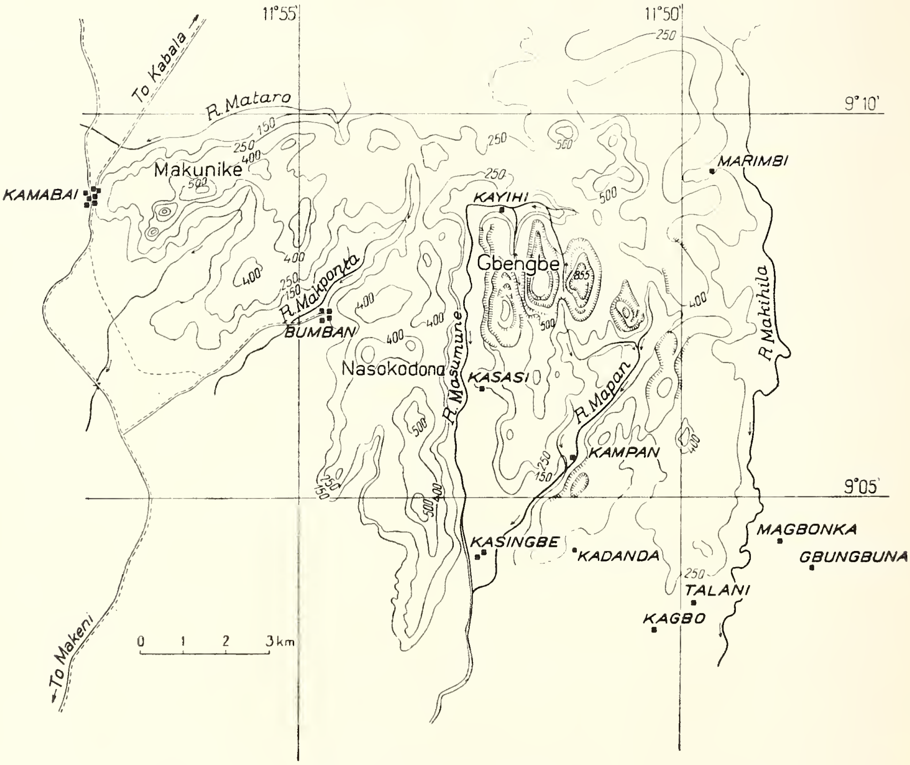

Fig. 2. The Gbengbe Hills.

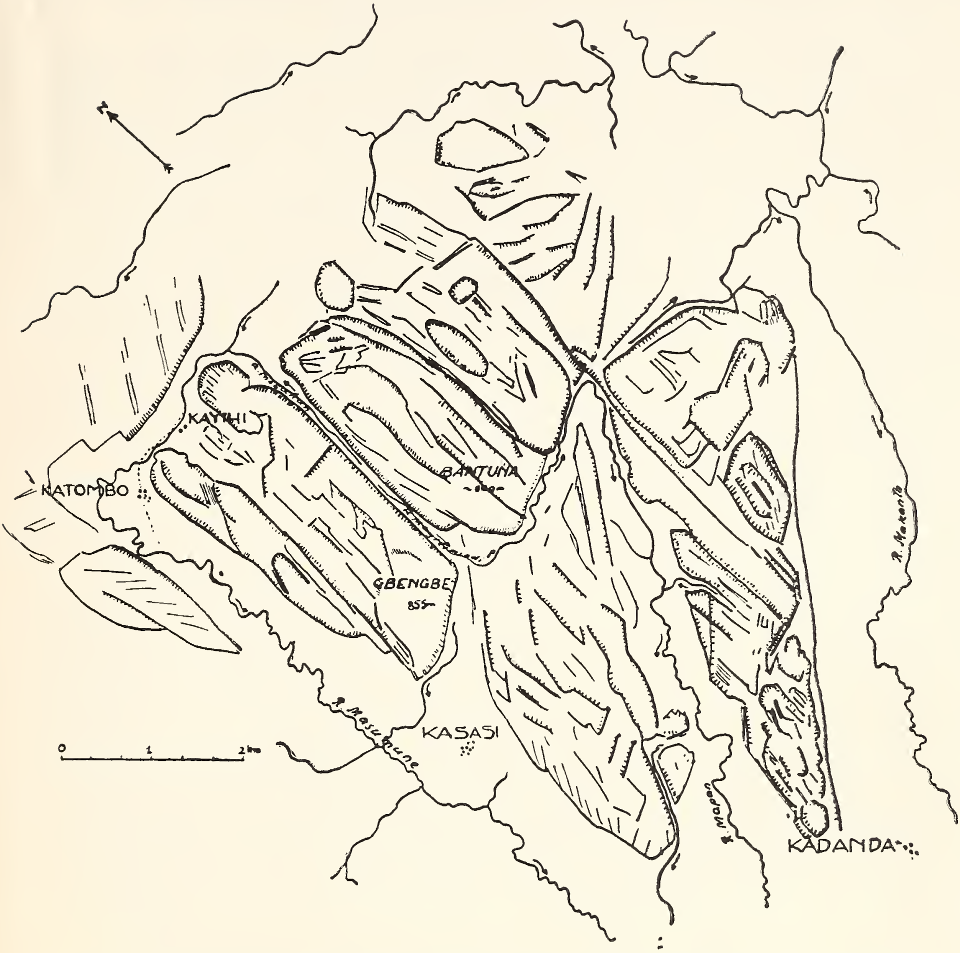

Fig. 3. Central part of the Gbengbe Hills. Prepared from the aerial photographs.

700 m above the flat environs in the West. The Gbengbe Hills are composed of porphyroblastic granite, and they include several hills, actually formed due to a rectangular splitting of a huge porphyroblastic granite block. The Gbengbe Hills are mostly barren excepting omnipresent grass cover; but even this is absent from the steep, smooth, black slopes of the rounded single hills. The highest peaks of the area are following: the Gbengbe (855 m); the Bantuna (appr. 800 m), separated by a deep narrow gorge from the Gbengbe; the Makunike (appr. 750 m), N of Bumban; and the Nasokodona ridge (appr. 550 m),

E of Bumban. The Gbengbe Peak elevates, like a steep black wall, about 650 m above the embracing small bush.

As mentioned above, the Gbengbe Hills compose a huge block, which is splitted rectangularly into several peaks. The main direction of such a splitting (Fig. 3) is N—S, which is also the main direction of fracturing and jointing of the porphyroblastic granite. The same direction is also that of rivers, which flow through the Gbengbe Hills, usually in stright valleys. The Masumune is the largest river of the area, and the valley in which it flows, separates the Nasokodona ridge from the central block of the Gbengbe Hills. In the same direction there is a valley near Bumban, on the western side of the hills, and another one runs along the eastern side of the hills, where the Makihila stream flows.

Another morphologically important direction of river valleys and also of fractures of the rock, is the SW—NE direction, represented within the area by the streams of Mataro, Makponta and Mapan (Fig. 2).

As mentioned valleys have a flat floor, mostly covered by soils, and they are intensively cultivated. Particularly the valley of the Masumune is suitable for such purposes, and there also cattle is conspicuously abundant.

The walls of mentioned valleys are always steep and high. The western wall of the Masumune valley elevates, in places almost vertical, on the average, within a horizontal distance of $^1/_2$ km, 200 to 250 m above the floor of the valley; the eastern wall elevates under similar conditions, south of Kasasi on average 100 m, north of Kasasi on average 550 m above the floor of the valley.

Along the valleys the villages of the Gbengbe Hills situate, and they are inhabited by people of the Limba tribe, who live on the cultivation of rice, kasawa, mallet, corn, etc.

THE GBOTOTO AND TENENE HILLS

The Gbototo and the Tenene Hills are actually names of parts of a coherent ridge, which occupies a much smaller area than do the Gbengbe Hills, and it situates south of Mabonto (Fig. 4). Between the Tenene and the Sula Mountains there is the Komboyan Hill (appr. 650 m), and between the Gbototo and the Sula Mountains is the Waka River. In the South, Gbototo dips down into the Pampana River, and in the West, there is a vast flat area (50 to 100 m a.s.l.), which is mainly covered by soils, probably alluvial. The highest top of the Gbototo rises 440 m, and that of the Tenene, appr. 400 m above sea level. Both hills together occupy an area of 20 km^2.

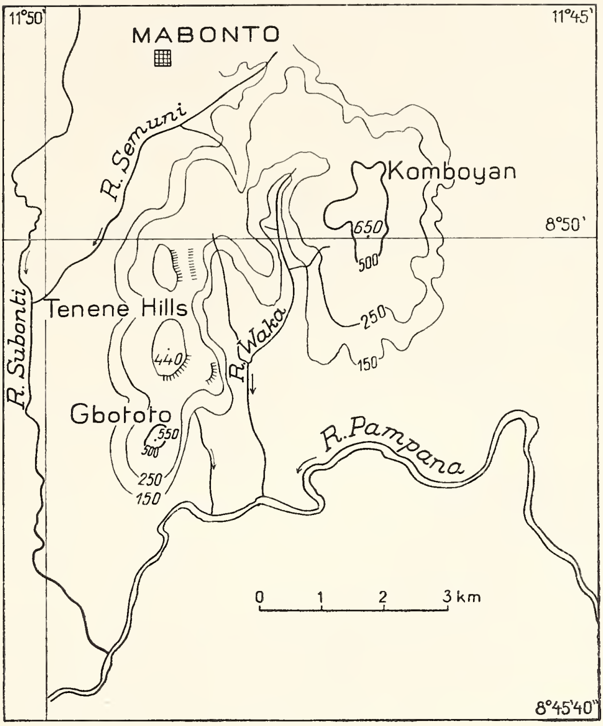

Fig. 4. The Gbototo, Tenene, and Komboyan.
All hills are composed of porphyroblastic granite.

The Gbototo and the Tenene are coming far behind the Gbengbe Hills, if their beauty is concerned; geologically and morphologically, however, they are attractive enough to be considered in the present paper.

GEOLOGY OF THE GBENGBE HILLS

Petrology

The geology of the Gbengbe Hills is rather simple, because the hills are composed of rather homogeneous porphyroblastic granite, and they are surrounded by an »aureole» of gneiss, often of migmatitic character. In the marginal parts of the mountains latekinematic aplite and granite veins and dykes occur.

The porphyroblastic granite contains large insets, upto 1 to 2 inch across, composed of microcline, oligoclase and quartz, and embedded in a groundmass consisting of quartz, oligoclase, and biotite, and of small amounts of microcline.

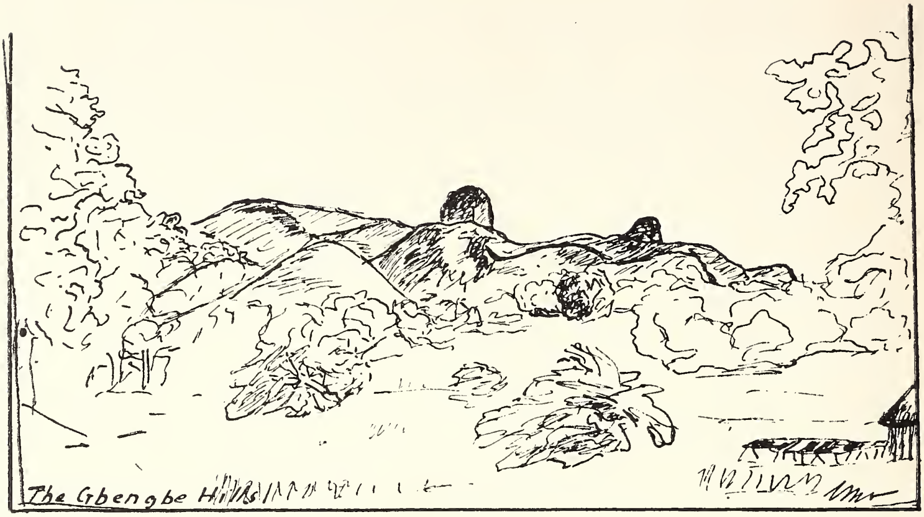

Fig. 5. The Gbengbe Hills as seen from the Southwest.

The last-mentioned mineral occurs in interstices of other constituents, or may be filling fissures and fractures of the rock. In spite of the granitic bulk composition of the rock, its matrix is definitely granodiorite. In places it may be even quartz diorite. In other words: the vast majority of the microcline of the porphyroblastic granite is concentrated into the porphyroblasts, the main potassium content of the matrix being fixed into biotite.

In the central part of the Gbengbe Hills, where the porphyroblastic granite is most homogeneous, the quartz veins occur, but they are rather sparse. They increase in amount towards the marginal part of the mountains, and they are rather abundant in the migmatitic gneiss aureole. Also the latekinematic aplite and granite veins are almost absent from the central parts of the mountains, but they are abundant in the marginal parts of the porphyroblastic granite, where they brecciate and migmatize the main rock.

The latekinematic granite is rich in potassium, contains muscovite and only a little biotite, sometimes also some epidote. The plagioclase is albitic, and the rock is medium- to fine-grained. Its microcline forms only occasionally interstitial filling, but it occurs equally to other constituents of the rock, which mode of occurrence of microcline is typical of the Sierra Leonean latekinematic granites, and it is different from that of the microcline of the synkinematic granites, where the microcline mostly is interstitial, or it occurs replacing the plagioclase.

The migmatitic gneiss, which surrounds the porphyroblastic granite hills, is mainly of granodioritic composition, and often strikingly poor in potash feldspar. Further away from the hills, the gneiss grades into more homogeneous rocks, into gneissose granodiorite, which may contain small portions of synkinematic granite.

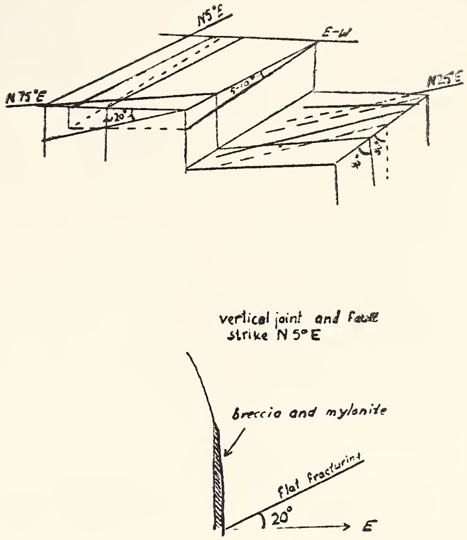

Fig. 6. Fracturing of the porphyroblastic granite of the Gbengbe Hills, and the structure of the southern part of the gorge between the peaks of Gbengbe and Bantuna.

As an example illustrating the relations between the described rocks composing the Gbengbe Hills, a geological section from SE towards NW will be described. In the Southeast, at Gbungbuna (Fig. 2) granodiorite gneiss occurs, and it contains minor strips of amphibolite. Between Gbungbuna and Magbonka the quartz veins increase in amount, and the gneiss, the strike of which is constantly N—S, grades in places into ader gneiss.

At Talani, porphyroblastic granite appears at the first time, but there it is still gneissose, and the latekinematic aplite veins are there abundant. Similar rock, accompanied by ample aplite veins, continues up to half way towards Kadanda; thence the aplite is rapidly decreasing, and at Kadanda the aplite veins are rather few, and do not occur at all northward of this locality.

In the northwest, between Kamabai and Bumban, the porphyroblastic granite verges on the migmatitic gneiss, which gradually obtains more and more aplitic metasome. At Kamabai, a beautiful pink aplitic granite, rich in potassium, outcrops. In this direction, the latekinematic granite enters also such portions of the porphyroblastic granite, which are only a little contaminated by foreign materials, as for instance along the valley of the stream between Bumban and Kamabai, consequently following the direction of a presumed shearzone.

In the topography, the hills composed of the porphyroblastic granite, clearly can be distinguished from those composed of gneiss, the latter being, in general, lower, and usually covered by small bush vegetation.

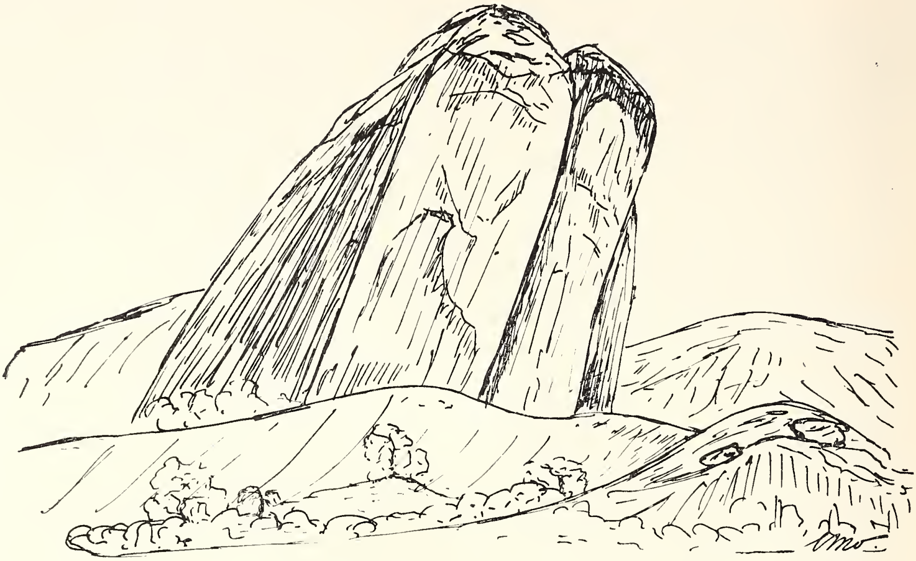

Fig. 7. A splitted peak of porphyroblastic granite.
The north-west end of the Gbengbe Peak, E of Katombo.

In the interior of the Gbengbe Hills, between the peaks of Gbengbe and Bantuna, there is a stright gorge in N—S direction. Along this gorge, distinct shearing occurs (Fig. 6), and in places there are thin portions of mylonite, composed of plagioclase and quartz, cemented by epidote, but there are no signs of any kind of potassium introduction, and even micas are absent, but sometimes chlorite occurs.

Structure

As one can see in Fig. 3, within the Gbengbe Hills, parallel N—S faults are characteristic. Between the big shearzones seen in Fig. 3, there are multitude of minor shear zones, wich are only few meters in width, but stretching N—S they may be rather extensive in length. Such shear zones are schistose and crushed, and the schistosity appears in a parallel arrangement of biotite flakes.

In addition there are distinct, but less frequent sheardirections E—W and approximately SW—NE. In Fig. 6 an example from the southern slope of the Bantuna is shown, the fracture system of the slope being illustrated. There is also seen a flat fracturing, the presence of which is rather typical of all porphyroblastic granites of Central Sierra Leone, and particularly this fracturing is responsible for the scaling of hills composed of porphyroblastic granite

(p. 11). The dip of the flat fracturing is usually 10° to 15° towards varying directions, but may frequently be as flat as 0° to 5°.

The mentioned shear zones are always present in the river valleys, as well. This fact may be the explanation for the origin of respective rivers, sheared portions being much more capable for disintegration and wearing down than are the portions composed of massif porphyroblastic granite. Consequently, the river valleys within the Gbengbe Hills reflect the zones of major shearing and faulting within the area under consideration. Similar zone is also the above-mentioned gorge between the Gbengbe and Bantuna. It has extremly steep walls (dip often 70 to 80°), and they rise in several places of the valley more than 50 m above the floor of the gorge, which is composed of strongly sheared porphyroblastic granite with minor mylonite portions, as described above. In this gorge, in the sheared floor, the microcline porphyroblasts have grown later than the shearing took place (Fig. 8).

In the middle of the gorge there is a windgap dividing two streams, one of which, the Mayilimindi, flows southwards, the other one, the Mabahun, flows northwards and joins at Kayihi the Masumune, which follows there a E—W valley. W of Kayihi, however, it turns southwards and follows then the large N—S valley mentioned above.

The E—W direction of faulting, particularly in the southern part of the Gbengbe Hills, is of minor importance; on the northern side of the line drawn from Bumban to Marimbi, this direction is, however, even more predominating than the south-northern one.

All mentioned directions, are usually well seen from the shape of the hills, as well, they often being splitted along the mentioned fracture plains. To the South of Katombo, on the eastern side of the Masumune valley (Fig. 7), there are high peaks with smooth, subvertical slopes stretching in the direction of the

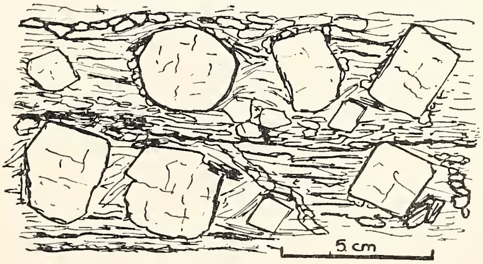

Fig. 8. Microcline porphyroblasts growing across a sheared portion of the matrix of the porphyroblastic granite. Southern slope of the Bantuna.

valley. Also east to Kamabi there are several hill tops which are splitted blockwise, and the direction of the splitting is that of the main jointing and fracturing.

THE GBOTOTO AND THE TENENE HILLS

In the areas around and south of Mabonto (Fig. 1), similar features as described above occur, as well, and best represented they are by the Gbototo and the Tenene Hills (Fig. 4).

The petrological structure of these hills is very the same as that of the Gbengbe Hills, with an exception: the latekinematic aplitic granite is there much more abundant than within the Gbengbe Hills, and there it brecciates not only the surrounding gneiss, but also the massif porphyroblastic granite itself.

In the South, near the Pampana River, the gneiss contains large, scattered insets of feldspars and quartz. Northwards, the amount of insets is rapidly increasing, and in the road cuts of the motorroad between Magburaka and Kalmaro, which is passing the Gbototo at its southernmost end, true »augen gneiss» with augens up to 5 cm across, occurs. Thence, the steep south slope of the Gbototo begins, entirely composed of homogeneous porphyroblastic granite. Further in the North, between the Gbototo and the Tenene Hills, latekinematic aplite granite forms abundant veins, up to 2 m wide, and their amount is rapidly increasing when entering the southern parts of the Tenene Hills. In the south central parts of the latter, the latekinematic aplitic granite may be so abundant, that it often appears as difficult to decide, if the rock should be mapped as latekinematic granite or migmatite.

Northwards the latekinematic granite is decreasing, and the north end of the Tenene Hills is composed of porphyroblastic granite only.

The Gbototo and the Tenene Hills are surrounded by gneiss, which is usually rather migmatitic. Particularly the Gbototo clearly differs in the topography from its gneissose environments. There the Gbototo, composed of the porphyroblastic granite, rises 300 to 400 m above the flat area on its western side, which is composed of gneiss. In the East, instead of a flat area, there are severall small gneiss hills, much lower than the Gbototo, and they are covered by bush, the Gbototo having a short grass cover.

Structurally, the Gbototo and the Tenene Hills are less clear than are the Gbengbe Hills. The same shear directions (N—S and E—W) occur there, as well, but they are of less extent being mainly restricted to directions of fracturing. The rivers on both sides of the Hills, the Subonti and the Waka, however, definitely follow a south-northern shear direction.

ON THE ORIGIN OF THE PORPHYROBLASTIC GRANITE

In the previous pages, the petrology and structure of the Gbengbe Hills, and of the Gbototo and Tenene, have been described in much more details, than it is actually necessary for understanding of their simply geology. It has been done so by an aim to give all necessary facts and details, essential for the porphyroblastic granites of Central Sierra Leone, and which are considered to be necessary for the forthcoming discussion dealing with an as difficult question as is that of the origin of the porphyroblastic granite and of the morphological features typical of the hills composed of such granite.

The starting point of such a discussion may be the fact, that there are large insets of feldspars and of quartz embedded in a matrix of granodioritic to quartz dioritic composition, and that the major part of its potassium content has been concentrated into porphyroblasts. The potassium of the matrix is mainly fixed into biotite and muscovite, the microcline being there compara-

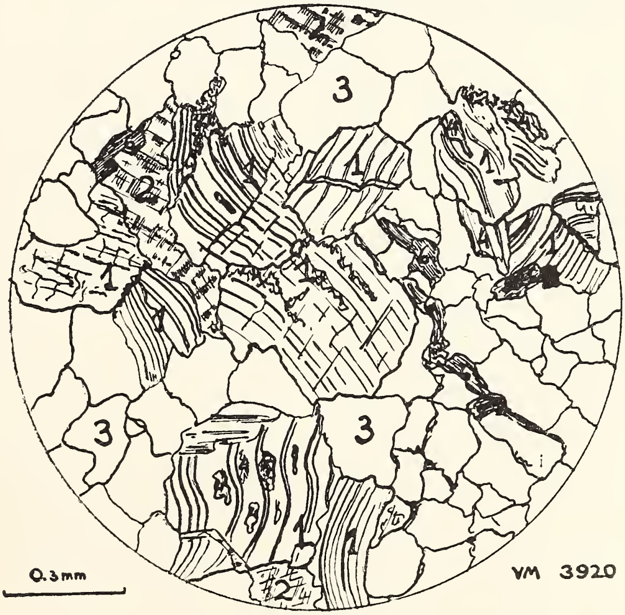

Fig. 9. Crushed portion of the porphyroblastic granite. Biotite (6) and plagioclase (1) lamellae are bent, and grains of quartz (3) are broken. In the plagioclase (oligoclase) there are small amounts of muscovite (4) and of epidote (5). Microcline (2) is interstitial and does not show any signs of shearing effects. The River Pampana, near Masimo. Crossed Nicols used for feldspars.

tively sparse, often almost absent, and it is always interstitial. In all cases studied it could be easily interpreted to be younger than any of other constituents of the matrix.

The second starting point of such a discussion is the observation, that the pophyroblasts, particularly those composed of microcline, are frequently filling the fractures, and often they have grown across them (Fig. 8). In such places, the rock may be strongly sheared but the porphyroblasts are free from any features indicating shearing. The same condition applies also to the interstitial microcline. The lamellae of plagioclase of the matrix may be intensively bent, but the microcline has been left intact by the shearing (Fig. 9).

As the third important argument of this discussion is the fact, that the potash feldspar of the prophyroblastic granite is exclusively microcline, which, according to the recent investigations, for instance of Goldsmith and Laves (1954), is not stable at the higher temperature, and, under hydrothermal conditions, it is definitely unstable at the temperature range above that of the amphibolite facies. In all conditions obtained by experiments, always orthoclase has been formed and never microcline.

Consequently:

1. The microcline was formed definitely later than the other constituents of the porphyroblastic granite, and also later than the main shearing and fracturing took place.

2. The emplacement of microcline (its crystallization) took place well below the melting temperature of the rocks or of its other constituents, and under such conditions, which, on the experimental basis, are sofar unknown. Marmo (1955 b) has supposed, that one of the requirements of the formation of microcline instead of orthoclase may be a slow accumulation of material. There are also other theories concerning the origin of microcline, and among them the most important one is that considering the microcline as being an alteration product of orthoclase (i.g. Laves and Goldmith, 1954 b), an opinion which, however, could not be proved by the field observations of the present author, and particularly if the microcline replacing the plagioclase or filling the interstices is concerned.

3. The problem regarding the origin of the porphyroblastic granite may be splitted into three questions: the origin of the matrix of the porphyroblastic granite; the formation of the porphyroblasts; the emplacement of the porphyroblastic granite.

The matrix of the porphyroblastic granite

In Table I, composition of some porphyroblastic granites, according to the determination on the integrating stage, are given.

Table I. Modal composition of four porphyroblastic granites and of their matrix, as determined on the I-stage.

	Plagioclase	Microcline	Quartz	Biotite	Muscovite	Rest
1. Porphyroblastic granite, bulk composition, Sangbaya	41.0	27.0	26.5	3.0	1.5	1.0
2. As above, matrix	49.0	1.0	36.0	10.5	0.5	3.0
3. Porphyroblastic granite, bulk composition, Gbengbe Hills	39.5	27.0	30.0	2.0	0.5	1.0
4. As above, matrix	50.0	1.5	35.0	7.0	1.5	5.0
5. Porphyroblastic granite, bulk composition, Gbototo	42.0	26.0	27.0	3.0	0.5	1.5
6. As above, matrix	49.0	1.5	37.0	9.5	1.0	2.0
7. Porphyroblastic granite, bulk composition, Keuruu, Finland	41.0	27.0	26.0	4.0	0.5	1.5
8. As above, matrix	50.0	1.2	36.0	9.0	0.5	3.3

In spite of the granitic bulk composition of the porphyroblastic granite, the matrix is of a granodioritic composition. Table II contains a comparison of the matrix of the porphyroblastic granite with the composition of a granodiorite from Central Sierra Leone, which both appear to be very similar.

Table II. Modal composition of two granodiorites of Sierra Leone and of matrix of two porphyroblastic granites, as determined on the I-stage.

	Plagioclase	Microcline	Quartz	Biotite	Muscovite	Rest
1. Biotite granodiorite, N of Kakonta	57.2	—	33.0	5.1	—	4.7
2. Biotite granodiorite, S of Kasokira	50.0	1.5	35.5	5.4	—	7.6
3. Anal. 2 of Table I	49.0	1.0	36.0	10.5	0.5	3.0
4. Anal. 4 of Table I	50.0	1.5	35.0	7.0	1.5	5.0

Elsewhere in Central Sierra Leone, and particularly at the south end of the Kangari Hills and at the north end of the Sula Mountains, there are gneisses grading into massif granodiorite, and these gneisses have been formed from ancient sediments. Particularly south of the Kangari Hills, the granodiorite often contains gneissose portions extremly rich in quartz, and these gneisses grade into true quartzites belonging to the Kangari schist belt. Furthermore, there are garnetiferous portions both in gneiss and in quartzites, and the sillimanite-bearing sediments within the Kangari Hills find their counterparts among the granodiorite gneisses in the sillimanite gneiss.

There porphyroblasts may occur, as well, and they appear at the margins of the schistbelt in the quartzite, also, hence forming a rock with microcline and quartz porphyroblasts embedded in a quartzite matrix. Consequently, there seem to be sufficient evidence to illustrate, that the matrix of a porphyroblastic granite may well have been formed on the expence of the sediments. Usually such an alteration of sedimentary rocks has been attributed to the synkinematic granitization, which phenomenon means such a change in the bulk composition of the rock, which, without fusion, leads towards the composition of granite, hence an alteration, closely linked with the introduction of potassium, the relative abundance of which is typical of the rocks termed granite. Most of rocks here under consideration, however, and including the matrix of the porphyroblastic granite, are quartz diorites or granodiorites, consequently they are such rocks, which do not contain any remarkable amount of potassium. Accordingly, such rocks hardly have been formed due to *granitization*, because, as one can see in Table III, the conversion of shales into granodiorite means a removal, and not an addition of potassium.

Table III. Comparison of the alkali contents of the synkinematic rocks with the respective contents in shales and in graywackes.

	K_2O	Na_2O	$\frac{K_2O}{Na_2O}$	Remarks
Shales of the Mississippi delta	2.30	1.51	1.52	According to Pettijohn (1949)
Average shales	3.24	1.30	2.49	—»—
Average graywacke	2.6	1.0	2.6	
Biotite granodiorite, Makera	1.87	5.54	0.34	
Biotite granodiorite, Matele	1.69	4.58	0.36	
Porphyroblastic granite, French Ivory Coast	2.65—3.55	3.85—5.35	0.65—0.7	

For that reason, the present author has supposed (Marmo, 1954) that the typical feature of the regional metamorphic reconstruction of the rocks is the granodioritization, and not the granitization, removal and not addition of potassium, an idea which has been expressed by Engel and Engel (1953), as well. The potassium released during the granodioritization may have been transported into places corresponding to a level of lower free energy, from compressed into tensional areas; in such circumstances the removed potassium would be expected to reappear along the shearzones and in the axial culminations, a fact which could be proved in Central Sierra Leone, where the potassic varieties of synkinematic rocks usually occupy such places. In accordance with the discussion above, the matrix of porphyroblastic granite may well be assumed to have formed, during the synkinematic metamorphism, due to the granodioritization of clay sediments.

The formation of porphyroblasts

The conditions under which potassium will be released from the sediments, are not clear, and in spite of many proposals based on thermodynamics, there still remain too many uncertain factors to be explained. The field observation, that the potassic varieties, in Central Sierra Leone, are mostly bound to the shattered zones, the granodiorites poor in potassium, on the contrary, mostly occurring in the compressed areas, may support an assumption, that the potassium has been squeezed out by the pressure; its migration may have taken place either due to ion migration through a solid state, or along the intergranular films. The field evidence also indicates, that in similar way, the silica may also be transported, and, to certain extent, sodium, as well. Such a migration from a compressed area towards a tensional one well explains the formation of the interstitial microcline of the granitized granodiorite. It may also explain the formation of aplitic veins, providing there an accumulation of aluminum, potassium, sodium, and silica favour the same locii; and due to hydrothermal conditions this material may act as an intrusive, in some instances even like a magma of an ideal granitic composition.

In this connexion, however, it may be of interest to note the experimental findings of Brown et allie (1953). According to their work, the interstitial material of a granite contains 2—6 % SiO_2, 20—32 % Al_2O_3, 11—40 % Fe_2O_3, 15—45 % CaO, and only a little alkalies. Hence the composition of the interstitial material does not support the ideas presented above, too well. The authors, however, have obtained their material from the rock by using dilute acid. The conditions under which the alkalies may release from the sediments during the synkinematic metamorphism, undoubtadly are of much more effective kind.

If particularly considering the formation of the porphyroblasts, there the potassium of a rock has been largely concentrated into these insets. The porphyroblasts are evenly distributed thorough the rock in such a manner, that the earlier opinions concerning the intrusive nature of such granites, and the phenocryst character of such insets, are well understandable. The potassium necessary for the formation of porphyroblasts has evidently derived from the rock itself, because otherwise the homogeneous distribution of the insets within a rock is difficult to explain. Ramberg (1952) explained such porphyroblasts to have formed due to a concretionary crystallization, and such an explanation is rather dependable concerning such porphyroblastic granites which form minor patches within evengrained granodiorites, or regarding the porphyroblasts occurring in quartzite (p. 18) or in amphibolite. The same explanation, however, seems to be the only possible one for the insets of the porphyroblastic granites like those considered in the present paper, and a concretionary enrichment also explains the formation of insets consisting of quartz or of plagioclase.

The emplacement of the porphyroblastic granite

The fact that the bodies of the porphyroblastic granite of such hills as the Gbengbe Hills, the Gbototo and the Tenene, is rounded, and that they are embraced by gneiss, will suggest their emplacement due to a doming. Such an emplacement may be considered to have taken place either due to less specific gravity than that of the overlaying rocks, as is the case regarding the salt domes, or the doming may be due to an enlargement of the volume of the doming material, if tectonic movements are, contemporaneously, opening the necessary space. The former alternative is out of question, because the specific gravity of the porphyroblastic granite and of the embracing gneiss is almost the same. The second alternative is worth of consideration.

The gneiss which surrounds the porphyroblastic granite bodies, often contains schistose portions of only slightly granodioritized metasediments, and the porphyroblasts are there more or less occasional or entirely absent. These facts point out, that the metamorphism of the gneiss could not take place under entirely similar conditions as under which the porphyroblastic granite was metamorphosed.

Because the latekinematic aplitic granite seems to accompany the porphyrpblastic granite, but occurs mostly along marginal parts of the body or along contacts between the granite and gneiss, there obviously exists some connexion between the intrusion of the latekinematic granite and the emplacement of the porphyroblastic granite.

Marmo (1955) has described a dome-like formation N of Bumbuna, about 15 km E of the Gbengbe Hills, where the doming material was latekinematic granite itself and there the shell of the dome consists of gneissose synkinematic granodiorte. In the western part of the dome-like structure, minor portion of the porpyroblastic granite occurs, as well. The necessary space for the doming was opened, at Bumbuna, due to N—S faulting, which produced a rotational movement and local folding, and the domed latekinematic granite occupied there the culmination of the formed fold.

In the Gbengbe Hills, the south-northern shearing is distinct and common, and probably it is related to the same movements which caused similar shearing and faulting at Bumbuna. In the Gbengbe Hills, this shearing preceeded (p. 11) the formation of porphyroblasts, and the movements in this direction may also there be attributed to the opening of the space necessary for the doming from below. Because the fractures and joints of the same direction have well survived the doming (or may have originated during the doming), the emplacement of the porphyroblastic granite obviously happened in a plastic solid condition. At the final stage of the doming, the potassium accumulated, due to a concretionary enrichment, into insets, because, if this accumulation should have preceeded the ceasing of movements related to the doming, the insets would hardly be as free from all shearing as they actually are. At the same stage, probably also the plagioclase and quartz insets were formed, sodium and silica having, at this stage, migrated towards appropriate crystallization centres, where they were able to deposit. Obviously the migration of the ions during the concretionary enrichment took place through the solid, and, consequently, was slow enough to enable the formation of microcline instead of orthoclase (see p. 16), this monocline modification of the potash feldspar being typical of the phenocrysts of true extrusive rocks of a rhyolitic composition.

That the porphyroblastic granite could attain such a plastic form which enabled it to dome, and its alkalies and silica to easily migrate, may be explained by assuming, that particular portions of the earth's crust got under such conditions, where a local increase in temperature was possible.

The assumption, that, under such circumstances, the potassium was enabled to migrate, explains, that considerable amount of it escaped and formed accumulations outside of the doming body. The material yielding the matrix of the porphyroblastic granite was primarily sediment (p. 18), and, consequently, it was rich in water, as well. Also its water must have had escaped, and if also the water was accumulated into places where the alkalies and silica were forming the materials of the later on intruding latekinematic aplitic granite, the water vapours obviously established there the hydrothermal conditions, under wich the aplitic granite could intrude the weak points around the doming body, the contacts between the porphyroblastic granite and the embracing gneiss, where it appeared in form of veins and dykes.

In the doming of the porphyroblastic granite, under conditions outlined above, the increase in volume of the doming material does not require any additional material, but it may be attained due to increased heat. Such a phenomenon has here been assumed mainly basing on the fact, that the recrystallization of the porphyroblastic granite seems to be more advanced than that of the embracing gneiss, in the former all signs of the primary schistosity being entirely destroyed, but are well preserved in portions of the latter. The doming-like dislocation was, naturally, directed from compressed towards the tensional places.

MORPHOLOGY AS RELATED TO THE GEOLOGY

In Fig. 3, the aerial photographs of the central part of the Gbengbe Hills have been used for the preparation of the attached map. There the drainage pattern is rather rectangular, and it is excellently portrayed by the fracture system, typical not only of the Gbengbe Hills, but of the whole Central Sierra Leone. In addition, there some rivers follow a diagonal fracture direction, which is less common but still typical shear direction of large areas including the Gbengbe Hills, the Sula Mountains and the Kangari Hills.

In the drainage pattern, no signs of any kind of thrust can be seen, but the fact, that between the Gbengbe and the Bantuna Peaks, in the gorge, mylonite occurs (p. 13), some kind of faulting there must have taken place, and most probably it has been vertical, in good agreement with the ideas regarding the emplacement of the porphyroblastic granite due to the doming. Consequently, the doming continued still after the rock became entirely rigid again, and at this stage of evolution of the dome, the breaking of the core body into blocks must have taken place. The formed blocks are the cores of the peaks of the Gbengbe Hills, the shattered fracture and fault zones having, later on, developed by disintegration of the rock into gorges and valleys of the recent rivers.

Also the peculiar form of the tor-like peaks may be explained on a geological basis.

A rock entirely recrystallized into a massif, homogeneous material, evidently is stronger than the surrounding gneiss, and hence also less capable for a decay than is the latter. Thus the disintegration of the granite is slower than that of the gneiss and it results in a formation of hills of rounded form and composed of the core of a dome. The actual form of these hills is due to their fracturing system. The two subvertical joint directions are responsible for the steep slopes of such hills. The flat fracturing of the porphyroblastic granite (p. 12) is resulting in the skaling of the hills. It is noteworthy, that the flat fracturing is not stright but it is slightly curving, and it mostly follows the surface of the top-cup of a tor-like hill; and such a flat fracturing is present in every porphyroblastic granite which the present author ever has come across.

In the literature dealing with the morphology of such rocks, often an opinion has been expressed, that such a fracturing and subsequent skaling are due to daily and seasonal fluctuations of temperature. Virtually, particularly in the circumstances of Sierra Leone, such a vibration in temperature and related expansion and contraction of the rock is far too small to cause any features like those appearing in the scaling of the hill tops, and producing rock plates up to 3 m thick, and even thicker.

There exists a geological explanation for the flat fracturing of the porphyroblastic granite. If starting from the emplacement of the porphyroblastic granite due to the doming, and supposing, that such a doming was a result of local elevation in temperature (p. 21), the upwards movement of the body is already capable to produce an orientation of the strength within the body, and particularly to fascilitate a jointing parallel to the upper surface of the doming body. In addition, obviously the doming core is rising into environment of lower temperature than is that of the doming material. Consequently there takes place a cooling, as well, and the gradient of the cooling is obviously almost vertical to the surface of the doming body. In such circumstances, the planes corresponding to the same temperature (within the body), expectadly are also the planes of easiest fracturing, and the cooling of the body will result in a tendency to split along these planes. Being subparallel to the outer surface of the core of a dome, these planes are concave, and, consequently, scaling of the hills along these planes will produce the semispheric shape of the tops of hills. This tendency has been preserved in the porphyroblastic granites which form the recent tor-like hills, and decay of these rocks, accordingly, follows these planes and also the subvertical jointing planes, which obviously have originated due to rise of the core and to general movements typical of the whole pre-Cambrian area of Central Sierra Leone.

The steepness of the slopes is a satisfactory explanation for the absence of vegetation from these slopes. That also the hill tops are devoid of any other vegetation but short grass, there must several factors be taken in account: the scaling of the rocks disturbes all perennial vegetation, particularly bush and trees. An absence of larger plants makes the high situated hill tops too exposed for wind to supply a suitable growing ground for larger annuals. In addition, the tops are, in general, dry, and, peculiarly, there are growing many grass species, which seem to be characteristic of the open and barren laterite duricrust surfaces of the Sula Mountains, as well.

CONCLUSIONS

From the evidence obtained from the study of the porphyroblastic granites of Central Sierra Leone, the following conclusions are reached:

The porphyroblastic granites are derived from sediments due to the granodioritization, which appears in the composition of the matrix of the porphyroblastic granites, it being poorer in potassium than are the sediments from which it has been supposed to have derived. The potassium, originally contained by this material, has been concentrated into porphyroblasts, which a concentration occurred due to concretionary enrichment.

The emplacement of the porphyroblastic granites has taken place due to a doming and owing to a joint action of pressure and of a local increase in the temperature there, from where the doming begun. The porphyroblastic granites domed in such places, where the space was opened due to the tectonic evolution of the area.

The fracturing and regular jointing of the porphyroblastic granite are the consequences of the main tectonic features of the large area, a part of which are the domed places. The vertical fracturing may also be a result of the upward movement of the porphyroblastic granite. The flat fracturing appears, mainly, owing to the cooling of rising body.

The concretionary crystallization of the porphyroblasts took place later than the fracturing of the matrix.

In addition to the formation of porphyroblasts, there evidently has been an excess in potassium, which was removed from the rock. After the doming was completed or nearly so, this released potassium together with likewise extracted sodium, aluminum and silica, formed a material which intruded the weak points around and within the domed body, and formed latekinematic aplitic granite veins, typical of the porphyroblastic granite areas described above.

Concerning the morphology of such, often tor-like hills composed of a porphyroblastic granite, in the case of the Gbengbe Hills, the rectangular drainage pattern clearly follows the vertical directions of the fracturing and shearing.

The scaling of such hills is supposed to be a result of the presence of the typical flat fracturing, which appears as a consequence of the upwards movement and subsequent cooling of the doming core composed of the porphyroblastic granite.

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